Department of Natural Resources
MARYLAND GEOLOGICAL SURVEY
Kenneth N. Weaver, Director

REPORT OF INVESTIGATIONS NO. 32

GEOLOGICAL INTERPRETATIONS OF AEROMAGNETIC MAPS OF THE CRYSTALLINE ROCKS IN THE APPALACHIANS, NORTHERN VIRGINIA TO NEW JERSEY

by

George W. Fisher, Michael W. Higgins and Isidore Zietz

1979

LIST OF ABBREVIATIONS AND TABLE FOR CONVERTING METRIC UNITS TO ENGLISH UNITS

MULTIPLY METRIC UNITS	BY	TO OBTAIN ENGLISH UNITS
Millimeters (mm)	0.03937	inches (in)
Centimeters (cm)	0.3937	inches (in)
Decimeters (dm)	0.3281	feet (ft)
Meters (m)	3.281	feet (ft)
Kilometers (km)	0.6214	miles (mi)
Square kilometers (km^2)	0.386	square miles (mi^2)

Department of Natural Resources
MARYLAND GEOLOGICAL SURVEY
Kenneth N. Weaver, Director

REPORT OF INVESTIGATIONS NO. 32

GEOLOGICAL INTERPRETATIONS OF AEROMAGNETIC MAPS OF THE CRYSTALLINE ROCKS IN THE APPALACHIANS, NORTHERN VIRGINIA TO NEW JERSEY

by

George W. Fisher[1], Michael W. Higgins[2] and Isidore Zietz[3]

1979

[1] Johns Hopkins University, Baltimore, Maryland
[2] U.S. Geological Survey, Reston, Virginia
[3] U.S. Geological Survey, Reston, Virginia

COMMISSION OF
MARYLAND GEOLOGICAL SURVEY

M. Gordon Wolman, *Chairman*.....Baltimore
S. James Campbell.....Towson
James M. Coffroth.....Frostburg
Richard W. Cooper.....Salisbury
John C. Geyer.....Baltimore

TABLE OF CONTENTS

	PAGE
Abstract	1
Introduction	3
The Aeromagnetic Map	3
Precambrian Crystalline Rocks	4
Reading Prong	4
Magnetic Character	4
Southwestern Reading Prong	5
Northeastern Reading Prong	5
Anticlines near Philadelphia and West Chester, Pennsylvania	10
Older Precambrian Rocks near Baltimore, Maryland	12
Magnetic Character	12
Geologic Interpretation	13
Conclusions	19
The Glenarm and Related Rocks	19
Magnetic Properties and Distribution of Rock Types	21
Setters Formation	21
Cockeysville Marble	21
Wissahickon Group	21
Stratigraphic Relations in the Glenarm	26
Sedimentary Facies Relations in the Glenarm	28
Interpretation of Sedimentary Environments	31
Age and Correlation of the Glenarm	33
Plate Tectonics and the Northern Piedmont	37
References Cited	41

Figures

		PAGE
Figure 1.	Comparison of observed and calculated magnetic anomalies for the southwestern Reading Prong	6
Figure 2.	Hypothetical tectonic map of the Reading Prong to illustrate the fold nappe hypothesis	8
Figure 3.	Cross section of Reading Prong	9
Figure 4.	Geologic sketch map of gneiss anticlines west of Philadelphia	11
Figure 5.	Directions of remanent magnetization for gneiss samples from the Towson anticline, Baltimore County, Maryland	17
Figure 6.	Directions of remanent magnetization for gneiss samples from the Woodstock anticline, Howard County, Maryland	17
Figure 7.	Directions of remanent magnetization for gneiss and quartzite samples from the Phoenix anticline, Baltimore County, Maryland	17
Figure 8.	Summary of directions of remanent magnetization for gneiss and quartzite samples from the Towson, Phoenix, and Woodstock anticlines, Baltimore and Howard Counties, Maryland	17

TABLE OF CONTENTS (Continued)

PAGE

Figure 9. Interpretative cross section of gneiss anticlines near Baltimore 18

Figure 10. Photograph of polished slab of albite-chlorite phyllite in Wissahickon 23

Figure 11. Laminated mica-quartz schist of Wissahickon 24

Figure 12. Metagraywacke interbedded with mica-quartz schist in the Wissahickon 25

Figure 13. Diamictite of the Wissahickon, showing clasts of metagraywacke in matrix of weakly foliated mica-quartz schist 27

Figure 14. Diamictite of the Wissahickon, showing clast of biotite-quartz gneiss in matrix of weakly foliated garnet-mica-quartz schist 28

Figure 15*a*. Distribution of major magnetic/lithologic sequences in the northern Piedmont 29

Figure 15*b*. Schematic tectonic map illustrating interpretation that the distribution of sequences B and C in the Wissahickon reflect interference of two fold systems 30

Figure 15*c*. Map showing generalized distribution of rock types in sequence C of Wissahickon 30

Figure 16. Restored stratigraphic cross section of the northern Piedmont from Baltimore, Maryland through Hanover to Carlisle, Pennsylvania 35

Figure 17. Sketch map illustrating the interpretation that the generalized shape of the carbonate bank edge could be inherited from the geometry of Late Precambrian rift systems that initiated the opening of the Early Paleozoic ocean east of North America 38

Figure 18. Plot of weight percent K_2O versus weight percent SiO_2 for northern Piedmont igneous rocks 39

Tables

Table 1. Magnetic data for samples from the Baltimore Gneiss and overlying Setters formation 14

Table 2. Stratigraphic nomenclature in the Glenarm Supergroup 20

Table 3. Magnetic susceptibilities of Piedmont rocks 22

Table 4. Modal analyses of felsic gneiss clasts from Wissahickon diamictite 32

TABLE OF CONTENTS (Continued)

Plates

Plate 1. Aeromagnetic map of eastern Pennsylvania and northern New Jersey showing generalized geologic relations of the crystalline rocks..... pocket

Plate 2. Aeromagnetic map of central Maryland, southern Pennsylvania, and Delaware showing generalized geologic relations of the crystalline rocks..... pocket

Plate 3. Generalized geologic map of the crystalline rocks in central Maryland, southern Pennsylvania, and Delaware..... pocket

Geological Interpretation of Aeromagnetic Maps of the Crystalline Rocks in the Central Appalachians, Northern Virginia to New Jersey

by

George W. Fisher, Michael W. Higgins, and Isidore Zietz

ABSTRACT

Previously published aeromagnetic data have been compiled at a scale of 1:250,000 for the northern Piedmont of the Appalachians by removing the International Geomagnetic Reference Field (E. B. Fabiano and N. W. Peddie, 1969, ESSA, Coast and Geodetic Tech. Rept. 38), selecting 100-λ contours. The resulting aeromagnetic map has been colored to portray field intensities more graphically; it clearly reflects many known geologic features, and provides valuable insight into several long-standing problems of Piedmont geology.

In the areas underlain by upper Precambrian (1 b.y.-old gneisses), the aeromagnetic maps provide information on the three-dimensional form of the contact between the gneisses and younger rocks. In the Reading Prong, the older Precambrian gneisses are associated with strong magnetic highs, and contrast markedly with adjacent Paleozoic sedimentary rocks, which are essentially nonmagnetic. The map patterns and profiles of the anomalies both support the interpretation that the Reading Prong is an extensive nappe that is refolded about northeast-plunging axes, and has its overturned limb exposed in synforms at the southwestern end of the Reading Prong and its upright limb exposed in antiforms in the northeastern Reading Prong.

The older Precambrian gneiss in the cores of the gneiss "domes" near Philadelphia and Baltimore is nonmagnetic and/or reversely magnetized, and is associated with deep magnetic lows. Well-defined anomalies associated with magnetic schists in the Wissahickon Group of W. P. Crowley (1976, Md. Geol. Survey Rept. Inv. 27) continue beneath the areas underlain by the older Precambrian rocks in several places, strongly supporting the interpretations of J. H. Mackin (1962, Geol. Soc. America Bull., v. 73, p 403-410) and Bromery (1968, unpub. Ph.D. dissertation, Johns Hopkins Univ.) that the Woodville and Avondale anticlines near Philadelphia, and the Phoenix anticline near Baltimore are refolded nappes, not simple domes.

The Glenarm Supergroup of Crowley (1976), which occupies a regional synclinorium between the South Mountain anticlinorium to the northwest and the anticlinorium cored by older Precambrian gneiss at Baltimore and Philadelphia, can be divided into three sequences on the basis of magnetic and lithologic properties: (A) an essentially nonmagnetic group of rocks including the Setters Formation and the Cockeysville Marble; (B) a group of weakly magnetic mica schists and phyllites, including the Ijamsville Phyllite, and the albite-chlorite phyllites and garnet-staurolite-mica schists of the Wissahickon; and (C) a strongly magnetic group of rocks including the diamictites, metagraywackes, and quartzose schists of the Wissahickon Group. The regional distribution of these three groups of rocks as inferred from our aeromagnetic data, supplemented by limited regional reconnaissance, and by detailed mapping of a few key areas, suggests that the three sequences approximate time-stratigraphic units. Sequence A can be traced almost continuously into the upper Precambrian to lower Paleozoic sequence of basal clastics and carbonate rocks, and appears to be the eastern continuation of these rocks. Sequence B appears to record submergence of the carbonate platform, and widespread deposition of argillaceous muds, presumably marine. Sequence C includes three distinct lithologies. On the southeast, it consists of a coarse, bouldery diamictite facies. The diamictite grades northwestward into a metagraywacke

facies, in which bedding thickness and grain size appear to decrease westward. Finally the metagraywacke appears to grade into thinly laminated quartz schists and phyllites. Together with rock compositions and relict sedimentary structures, these relations suggest that the entire sequence represents a single coherent package of flysch sediments, fed into the basin from the southeast. Fragments of volcanic and ultramafic rocks in the diamictite suggest that the source of the sediments was composed of rocks like those presently exposed in the Baltimore Complex (Baltimore Mafic Complex of Crowley (1976), which occurs in a thrust block just southeast of the Wissahickon diamictite. We therefore concur with Crowley's view that this thrust was emplaced during sedimentation, and we conclude that its emplacement may well have triggered deposition of the flysch sequence. The age of the flysch sequence is constrained in two ways: (1) it must be younger than the Baltimore Complex, which includes volcanic rocks of the James Run Formation, dated at approximately 550 m.y. by G. R. Tilton and others (1970, p. 429-437, *in* Fisher, G. W. and others, eds., Studies in Appalachian Geology; Central and Southern; New York, Interscience); and (2) it must be older than the Upper Ordovician Arvonia Slate (Middle to Upper Ordovician) of Virginia. Consequently, the entire Wissahickon Group may be in normal stratigraphic position on the Setters Formation and Cockeysville Marble, which appear to correlate with the lower Paleozoic clastic/carbonate section of the Blue Ridge.

The present distribution of carbonate rocks implies that a deep indentation may have existed in the edge of the carbonate bank northwest of Baltimore, as proposed earlier by John Rodgers (1968, p. 141-149, *in* Zen and others, eds., Studies of Appalachian Geology; Northern and Maritime; New York, Interscience). This configuration, the major bend in the Appalachians between Baltimore and New York, and the regional variation in thickness of the basal Cambrian sedimentary sequence could all reflect the original geometry of the facture system along which the Paleozoic ocean opened. One possible reconstruction is that the northern and southern Appalachians originated along fracture systems emanating from different triple junctions, and that the central Appalachians represent the area where the two rift systems were joined by a major east-west cross fracture, possibly a transform fault.

INTRODUCTION

With only a little poetic license, the Piedmont of the Appalachians can be characterized as one of the last frontiers of geologic study in the conterminous United States. Poorly known because of its complexity, poor exposure, and the low topographic relief, the geology of this area is nevertheless vitally important both to an understanding of the evolution of the Appalachians and as a data base for future urban development.

The northern Piedmont, from northern Virginia to New Jersey, is probably the best known part of the Piedmont, but even here many fundamental questions remain. Regional aeromagnetic maps of this area, recently compiled by the U.S. Geological Survey (Plates 1 and 2), reflect many features of the geology shown on published state geologic maps. For example, the shape of the basement gneiss anticlines near Baltimore, the form of the South Mountain anticline, and the outcrop pattern of diabase sills intruding the Triassic and Jurassic sedimentary rocks of the Newark Group are all clearly shown on the aeromagnetic maps. Many other points of correspondence have been discussed by Bromery and Griscom (1967), Bromery (1968), and others. Because of this correspondence, it seems obvious that interpretation of aeromagnetic maps can and should play a vital part in working out the geology of the Piedmont, especially in view of the speed and low cost of aeromagnetic mapping and in view of the information magnetic data can give on the third dimension of large structures, so difficult to determine by surface mapping in this area of low relief.

In this paper, we attempt to use aeromagnetic maps to aid in interpreting some aspects of the geology of the northern Piedmont that are particularly interesting to us. Our goals have been (1) to cast some light on important, selected geologic problems; (2) to work out the relation between magnetic properties and stratigraphic units in this relatively well known part of the Piedmont as a prelude to work in other, less known parts; and (3) to illustrate the usefulness of combining aeromagnetic interpretation with geologic mapping. As our work has progressed, we have become more and more convinced of the importance of using these approaches in conjunction with one another. In order to extract the maximum amount of useful information from aeromagnetic maps, it is vital to approach them from a geologic point of view, with a clear understanding of what the geologic problems are. Used in this way, aeromagnetic maps provide an invaluable "common denominator" in integrating on a regional scale the detailed mapping of local areas, often done by different people using vastly different approaches; in relating stratigraphic units defined by use of different criteria; and in tracing units through unmapped areas.

Many of our colleagues have aided in many ways during the slow evolution of ideas which led to this paper. David Elliott, Clifford A. Hopson, John Rodgers, and William Crowley deserve special mention, but many others have given freely of their time and ideas. We thank them all. Richard Goldsmith, Avery Drake, and Andrew Griscom have reviewed the manuscript and made many valuable suggestions, for which we are very grateful. The cost of publishing the color maps was defrayed by National Science Foundation Grant No. EAR 77-09968 to the Johns Hopkins University.

THE AEROMAGNETIC MAP

The aeromagnetic map (Plates 1 and 2) was compiled from published data indicated on the index map located on the margin of Plate 1. The flight line elevation and spacing for each area are described in "Sources of Aeromagnetic Data." Aeromagnetic data were obtained by means of an ASQ-3A fluxgate magnetometer coupled to compensating and continuous recording equipment. Standard navigational equipment and county road maps were used for guidance and the flight path was recorded by means of a continuous strip-film camera. Altitude was maintained by standard navigational equipment and by radar altimeter.

The aeromagnetic map (Plates 1 and 2) was derived by reducing the original maps when appropriate, removing the International Geomagnetic Reference Field (IGRF) (Fabiano and

Peddie, 1969), and selecting the 100-gamma contours to produce the colored map.

Coloring of the map is significant, for it helps to distinguish the intensities of the field, a feat which is almost impossible using black and white reproduction. In a sense, the map may be compared to a three-dimensional view of the field, the color adding a vertical dimension to a "two-dimensional" map. Finally, the colored version helps us to separate clearly the short wavelength from the longer wavelength anomalies, so that mathematical filtering is not only unnecessary, but undesirable.

PRECAMBRIAN CRYSTALLINE ROCKS

The structural relations between the older Precambrian (about 1 b.y-old) crystalline rocks and the overlying upper Precambrian and Paleozoic rocks of the Appalachians have long been controversial. Major obstacles to interpreting the three-dimensional form of the structures involved have been the poor exposures and lack of topographic relief. Fortunately the older Precambrian rocks have a magnetic character different from that of the younger rocks in most of the area shown on Plates 1 and 2, and therefore the aeromagnetic data provide important evidence bearing on the structures.

Within the area shown on Plates 1 and 2, the magnetic expression of the older Precambrian rocks varies markedly from north to south. North of an east-west line at approximately latitude 40°15′N, most of the older Precambrian rocks are highly magnetic, and are represented by the highest magnetic anomalies on Plate 1, as high as 300γ. The magnetic character of these rocks clearly reflects their high magnetite content (Drake, 1969). Farther south the older Precambrian rocks are less magnetic; between latitude 40°15′N and 39°50′N, the rocks are weakly magnetic, characterized by anomalies in the range of 600γ to 1000γ. South of 39°50′N, they are essentially nonmagnetic and are marked by the deepest lows on the magnetic map. These variations hint at regional differences in the character of the older Precambrian rocks and suggest that a study of this variation on a regional scale might help to delineate major magnetic and/or petrologic provinces within the older Precambrian rocks of eastern North America.

READING PRONG

The older Precambrian rocks within the area shown on Plate 1 are exposed in an elongate group of anticlines collectively known as the Reading Prong. The structural relations between the Precambrian rocks and the Paleozoic sedimentary rocks have been the subject of continuing controversy. Many authors have interpreted the anticlines as normal upright antiforms in which the Precambrian rocks are structurally beneath the Paleozoic rocks, but others have interpreted the Precambrian rocks as overlying the younger rocks, either in thrust slices or in synforms. The history of this controversy has been summarized ably by Drake (1970, p. 279-280) and Dallmeyer (1974, p. 132) and will not be repeated here.

Magnetic Character

Throughout the Reading Prong, the older Precambrian rocks have a high magnetic susceptibility; they are marked by irregular anomalies as high as 3000γ in amplitude, which outline the outcrop areas of the Precambrian rocks almost perfectly. These high magnetic susceptibilities clearly reflect the high magnetite content of most of these rocks (see modal analyses in Drake, 1969). Locally, variation in the intensity of the magnetic anomalies appears to correlate with variation in rock type. In northeasternmost New Jersey, for example, the highest anomalies are underlain by rocks shown on the state geologic map by Johnson

(1950) as "Losee Gneiss", whereas lower anomalies are underlain by rocks shown as Johnson's "Byram Gneiss". In southwestern New Jersey, however, the relative magnetic expressions of these two units are just reversed, indicating that the magnetic maps cannot be used alone to map the conventional lithologic units in the Reading Prong.

Southwestern Reading Prong

The structure of the Reading Prong has been acrimoniously debated for decades. Drake (1969, 1970) reviewed the complex evolution of thought on these rocks and made a strong case for interpreting the older Precambrian rocks of the southwestern Reading Prong as allochthonous masses above the younger Cambrian and Ordovician sedimentary rocks; his interpretation was based on (1) field evidence that the Precambrian rocks overlie the Cambrian and Ordovician rocks at several localities (Drake, 1969; 1970, and references cited therein); (2) regional gravity and seismic evidence that true basement is 9 to 14 km below the surface just west of the Reading Prong (R. N. Bromery, *in* U.S. Geological Survey, 1966, p. A70; and G. H. Wood and M. D. Carter cited *in* Drake, 1970); and (3) the pattern of aeromagnetic anomalies, which generally terminate abruptly at the northeastern and southwestern ends of areas underlain by older Precambrian rocks, instead of plunging beneath the Paleozoic rocks (Bromery and Griscom, 1967). This relation is particularly well shown at the southwesternmost end of the Reading Prong, at Reading, Pennsylvania; in the belt 15 km southwest of Bethlehem, Pennsylvania, in the area just south of Phillipsburg, New Jersey; and in the area just south of Washington, New Jersey. In each area the high magnetic anomalies are confined to the actual outcrop belts of older Precambrian rocks and only locally extend beneath the Paleozoic rocks. Drake (1969, 1970) considered, but rejected the idea that the older Precambrian rocks occur in simple imbricate thrust slices because, although low angle faulting is widespread just above the older Precambrian-Paleozoic contact (Drake, 1969, Fig. 2), the faults have produced only minor disruption of the stratigraphic sequence. Work in the Paleozoic sedimentary rocks just to the west had already revealed extensive nappe structures having sub-horizontal axial planes. Accordingly, Drake suggested that the older Precambrian rocks in the southwestern Reading Prong occupied synforms formed by refolding the lower limb of an extensive nappe, which he named the Musconetcong Nappe (Drake, 1970, Fig. 5).

As a test of Drake's interpretation, we have compared the magnetic anomalies observed along the line of Drake's section (Drake, 1970, Fig. 5) with the anomalies that should result if the Precambrian rocks occupy the cores of synforms (Fig. 1*a*) or antiforms (Fig. 1*b*); the predicted anomaly patterns were calculated graphically by use of a Pirson polar chart (Pirson, 1940). The two structural interpretations produce anomaly patterns that are broadly similar, but the details of the pattern predicted by Drake's synformal interpretation are in much closer agreement with the observed anomalies. The differences are most pronounced in the area southeast of Scott's Mountain, where the antiformal interpretation requires the magnetic highs developed over the Precambrian rocks to extend much too far southeast into the valleys occupied by the Paleozoic rocks, sharply pinching the lows over the Paleozoic rocks. The only areas where the synformal interpretation predicts a pattern markedly different from the observed pattern are along the northwest edge of the Scott's Mountain structure and at the southeastern edge of the Musconetcong Mountain structure. The anomaly pattern at Scott's Mountain suggests that the Precambrian rocks along the northwestern edge of the Precambrian belt here are thicker than shown in Drake's cross section (Drake, 1970, Fig. 5), but not as thick as in the antiformal cross section (this report, Fig. 1*b*). The agreement at the southeastern end of the cross section would be improved if a thin belt of weakly magnetic rocks were located near the center of the Musconetcong Mountain structure, and if the Precambrian rocks extend farther to the southeast, under the Triassic and Jurassic rocks than shown in Drake's cross section (Drake, 1970, Fig. 5). But in view of the generally good agreement between the anomaly pattern predicted by the synformal model and the observed pattern, we conclude that Drake's synformal interpretation better approaches the true structure than an antiformal interpretation.

Northeastern Reading Prong

As discussed by Drake (1969, 1970) and Dallmeyer (1974), the structure of the northeastern end of the Reading Prong appears to differ from that of the southwestern part. The distinctive magnetic pattern of the southwestern Reading Prong, in which most anomalies terminate abruptly at the end of the older Precambrian outcrop belts, does not continue into the northeastern

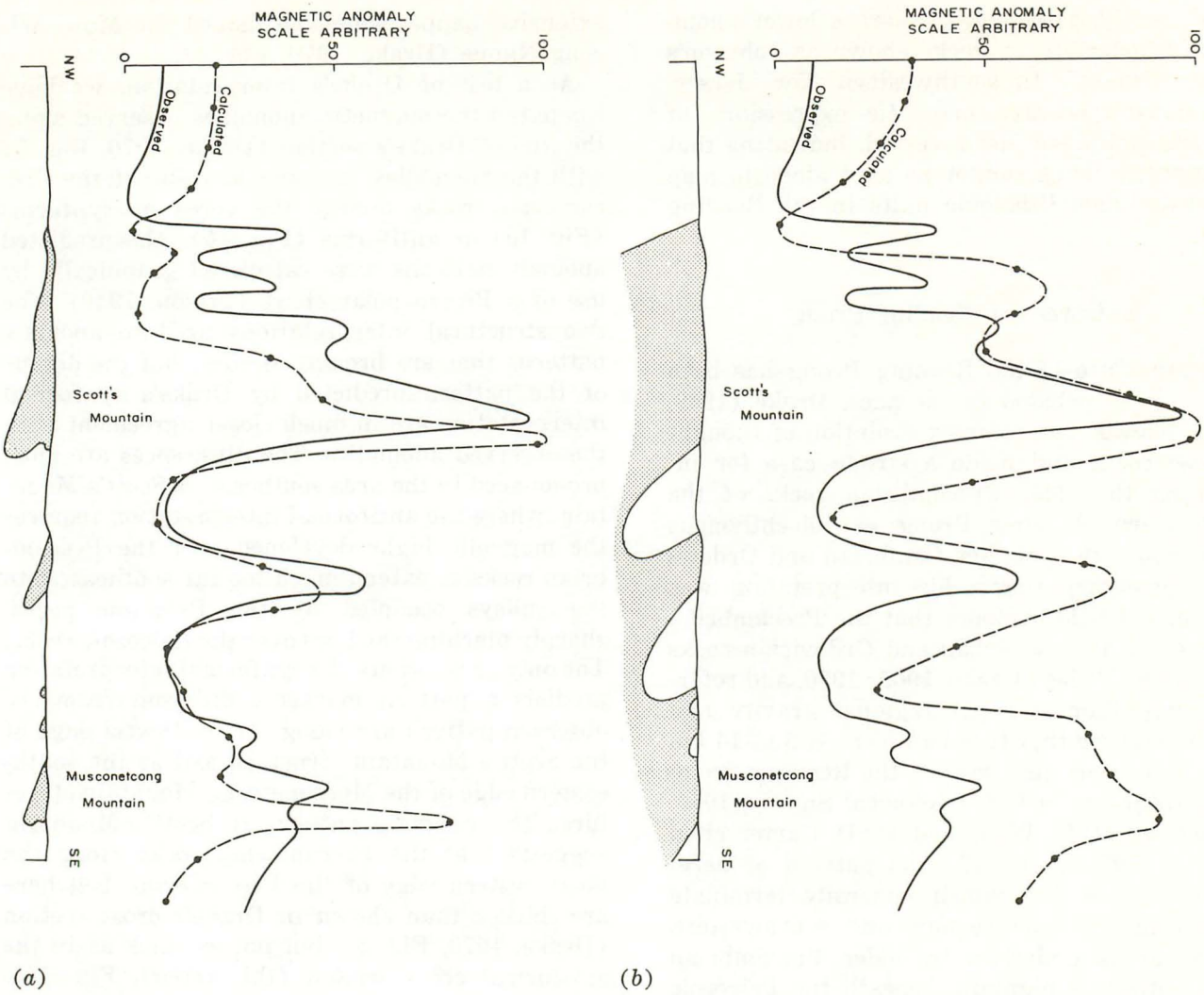


Figure 1. Comparison of observed (solid lines) and calculated magnetic anomalies (dashed lines) for the southwestern Reading Prong. Figure 1*a* assumes that the magnetic Precambrian rocks (stippled) occupy synforms, as proposed by Drake (1970, Fig. 5); Figure 1*b* assumes that the Precambrian rocks occupy antiforms. The line of section is the same as in Drake's (1970) Figure 5, and is shown on Plate 1. The overall amplitudes of calculated anomalies have been arbitrarily adjusted to match observed anomaly amplitudes.

Reading Prong. Instead, at several places, the anomalies associated with the older Precambrian rocks appear to continue beneath the Paleozoic rocks, suggesting that older Precambrian rocks occupy the cores of plunging antiforms (Plate 1). This can be seen clearly in the area around New Milford, New York; in the area 12 km southwest of Harriman, New York; and in the Green Pond synform between Middle Valley and Oak Ridge, New Jersey (Plate 1). A structure near Oak Ridge is especially convincing; just southeast of Oak Ridge, an elongate, northeast-trending anomaly with gentle gradients coincides almost perfectly with the anticline outlined by the contact between Silurian and Devonian rocks, suggesting that this anticline is cored by a buried mass of magnetic older Precambrian rocks. Similar features are widespread in the northeastern Reading Prong and suggest to Drake, to Dallmeyer, and to us, that the older Precambrian rocks in the northeastern Reading Prong occupy northeast-plunging antiforms.

In order to reconcile the apparent antiformal character of the older Precambrian rocks in the northeastern Reading Prong with the synformal character of the southwestern end, Drake (1969, 1970) suggested that the two parts of the Reading Prong could represent different limbs of the Musconetcong Nappe; the southwestern part could represent synforms on the lower, overturned limb,

and the northeastern part could represent antiforms developed on the upright limb. In this sense, Drake proposed that most of the Reading Prong is allochthonous.

Dallmeyer (1974) argued against the nappe interpretation, claiming that the aeromagnetic evidence already discussed and geologic evidence indicate that the northeastern Reading Prong is not "allochthonous". In assessing this argument, we must consider carefully what Drake and Dallmeyer mean by the term "allochthonous". The evidence cited by Dallmeyer (1974) clearly indicates that the rocks of the northeastern Reading Prong are predominantly right side up, that the older Precambrian rocks are exposed in antiforms having depths of at least a few kilometers rather than in thrust slices, and that the exposed older Precambrian rocks have not been extensively displaced laterally with respect to the nearby Paleozoic rocks; in this sense, the older Precambrian rocks are clearly autochthonous. But none of Dallmeyer's evidence conflicts with the interpretation of the northeastern Reading Prong as the upper limb of a major nappe cored by a slab of older Precambrian rocks possibly several kilometers thick, overlying still deeper Paleozoic rocks in a recumbent syncline.

As an aid to evaluating this program, we have attempted to interpret the outcrop of the Reading Prong in New Jersey as a nappe structure of the sort probably visualized by Drake; Figure 2 is a hypothetical tectonic map of the area, and Figure 3*a* is a corresponding cross section, produced by viewing the structure down the axes of the minor folds, most of which plunge gently northeast (Drake, 1969, p. 97). We can not attempt to evaluate the nappe interpretation by comparing the subsurface extent of the older Precambrian rocks required by the structures postulated in Figure 2 and 3*a* with that indicated by the aeromagnetic data. The critical feature of the nappe interpretation is that the older Precambrian rocks should continue beneath the Paleozoic rocks where synformal and antiformal axes leave the outcrop area of the older Precambrian rocks toward the northeast, and should be absent beneath the Paleozoic rocks wherever the same axes pass out of the area of Precambrian rocks in a southwesterly direction.

Plate 1 shows that these criteria are fulfilled in most areas. As already mentioned, the Precambrian rocks clearly plunge beneath the Paleozoic section at the northern end of the Reading Prong between New Milford and Harriman, New York. At the southern end of the Reading Prong, most high magnetic anomalies terminate abruptly at the ends of the outcrop belts of Precambrian rocks, as discussed in detail by Bromery and Griscom (1967). The small outlier of Precambrian rocks just northwest of Great Meadows, New Jersey, has precisely the expected configuration; the anomalies associated with the Precambrian rocks plunge beneath the Paleozoic cover at the northeast end of the Precambrian belt, whereas the southwestern end has virtually no magnetic expression and must be very thin if it is underlain by rocks like those at the northeastern end. Similarly, the Precambrian rocks just southeast of Washington, New Jersey, seem to be very thin.

Perhaps the most convincing feature of Plate 1 is the contrast between the magnetic pattern associated with the Paleozoic rocks in the Green Pond synform and that associated with the belt of Paleozoic rocks which parallels the Green Pond synform about 7 km to the northwest, and 5 km east of Washington, New Jersey. At the southern end of the Green Pond synform, near Middle Valley, New Jersey, the Paleozoic rocks clearly form a thin cover over highly magnetic Precambrian rocks plunging to the north, confirming the synformal structure. By contrast, the belt of Paleozoic rocks 7 km to the northwest shows no sign of magnetic rocks in the subsurface, even at its northern end, suggesting that it is a north-plunging antiform. The only major exception to the predicted pattern is in the area four kilometers east of Phillipsburg, New Jersey, where the magnetic pattern associated with the older Precambrian rocks extends about 2 km southwest into the area underlain by Paleozoic rocks in what ought to be a northeast-plunging synform according to Figure 2 and 3*a*. This area is, however, extensively covered by glacial drift (Johnson, 1950) and the Precambrian rocks may be more extensive than as mapped; or the complex faulting in this area may have disrupted the subsurface structure more than is recognized in Figures 2 and 3*a*. But in any case, the anomaly patterns throughout most of the Reading Prong are compatible with the nappe structure envisioned in Figures 2 and 3*a*.

On balance, then, the aeromagnetic data appear to support Drake's (1969, 1970) interpretation of the Reading Prong as a refolded nappe plunging gently northward. However, one varient of the nappe interpretation deserves serious consideration. The cross section shown in Figure *3a* assumes that the nappe is more or less cylindrical and that fold axes associated with the nappe parallel the regional strike. But if the fold is not cylindrical, the amplitude of the nappe could decrease along strike to the northeast, from a structure like that shown in Figure 3*a* in the southwestern Reading Prong to one like that shown in Figure 3*b* in the

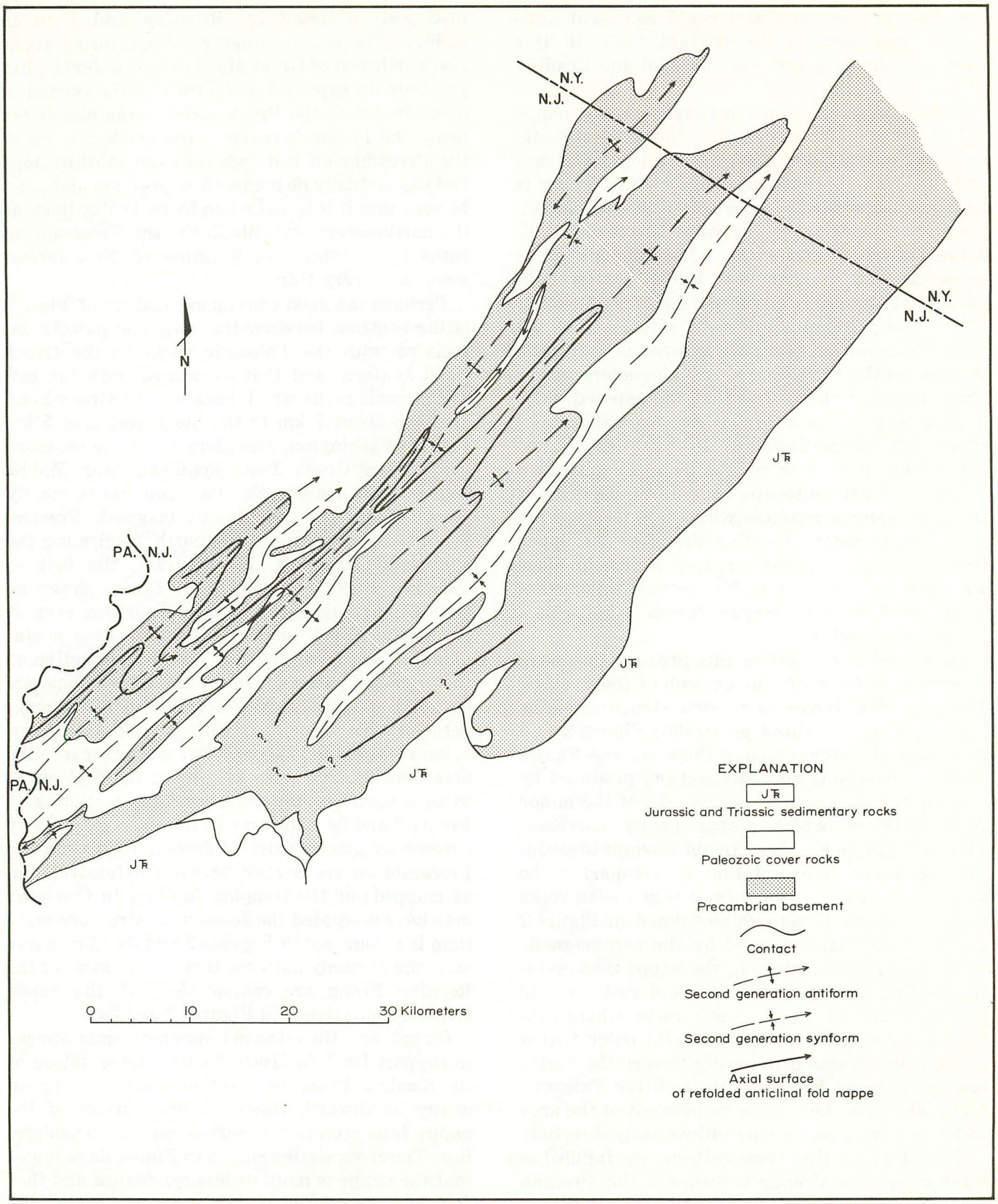

N.Y.
N.J.
N.Y.
N.J.
N
PA.
N.J.
PA.
N.J.
JTR
JTR
JTR
JTR
0
10
20
30 Kilometers
EXPLANATION
JTR
Jurassic and Triassic sedimentary rocks
Paleozoic cover rocks
Precambrian basement
Contact
Second generation antiform
Second generation synform
Axial surface
of refolded anticlinal fold nappe

Figure 2

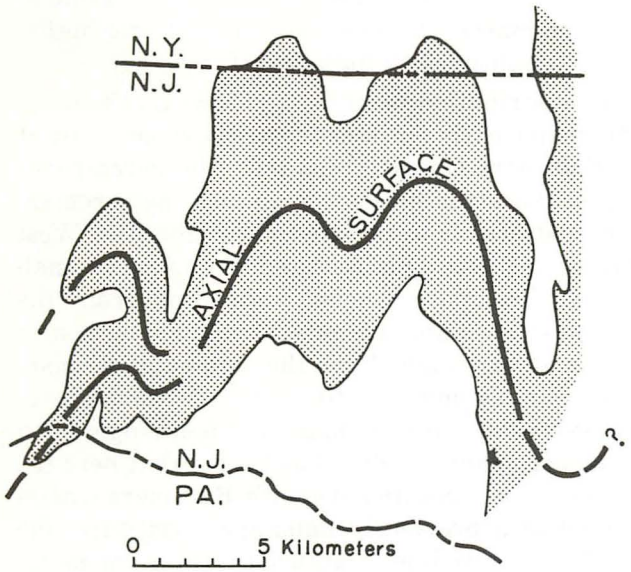


Figure 3*a*

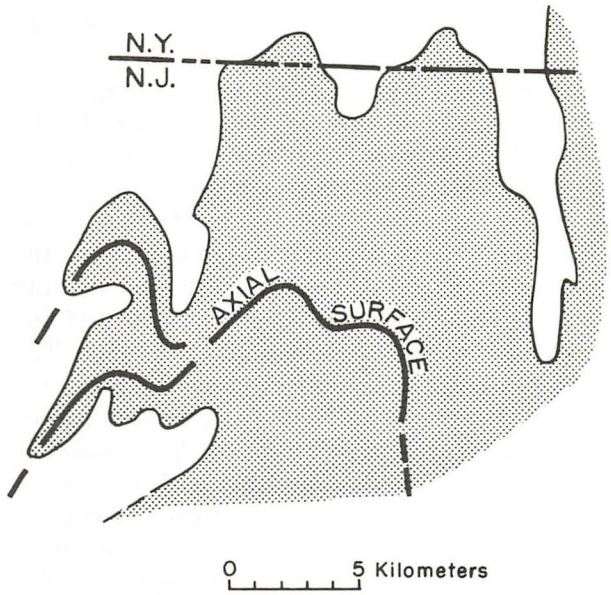


Figure 3*b*

Figure 3. Cross section of Reading Prong, constructed by viewing Figure 2 down the plunge of minor folds trending parallel to the regional strike, and plunging 7° northeast. Figure 3*a* assumes cylindrical folding, in which axes of early folds are parallel to those of later folds; Figure 3*b* is an alternative interpretation for the structure near the New York-New Jersey boundary, assuming that axes of the early folds trend more northwesterly than those of later folds. See text for discussion.

Figure 2. Hypothetical tectonic map of the Reading Prong to illustrate the fold nappe hypothesis; see text for discussion. The extent of the Precambrian rocks is modified from the State geologic maps for Pennsylvania (Gray and others, 1960), New Jersey (Johnson, 1950), and New York (Broughton and others, 1962).

northeastern Reading Prong. If the fold is not cylindrical, minor folds associated with the lower limb of the nappe should trend north of the regional strike. The aeromagnetic data provide one line of evidence favoring this possibility; as pointed out by Bromery (1960), the broad, low amplitude anomaly 8 km northwest of Bethlehem, Pennsylvania, suggests the presence of a deeply buried antiform of Precambrian rocks. This anomaly and the inferred antiform trend approximately 25° north of the regional strike. If the deep-seated structures parallel this anticline, the amplitude of the nappe may decrease to the north.

One major problem inherent in the nappe interpretation is the location of the root zone. In the interpretation sketched in Figure 2 and 3, only the part of the Reading Prong northwest of the Green Pond synform is clearly allochthonous. The rocks southeast of this synform and their equivalents to the northeast in New York may be rooted and may constitute the root zone of the nappe. The aeromagnetic data are compatible with this interpretation. The easternmost synform shown in Drake's cross section of the southwestern Reading Prong (our Fig. 1*a*) is part of the belt of older Precambrian rocks southeast of the Green Pond synform and is one place in which the magnetic anomaly calculated for Drake's cross section deviates substantially from the observed anomaly; as already noted, a better fit would be obtained by continuing the older Precambrian rocks southeastward under the Triassic and Jurassic rocks. The same conclusion is suggested by the pattern of anomalies in Plate 1, which imply that the magnetic older Precambrian rocks extend beneath the Triassic and Jurassic rocks to the southeast.

Anticline Near Philadelphia and West Chester, Pennsylvania

Just west of Philadelphia, another group of older Precambrian rocks (named the Baltimore Gneiss, after similar rocks near Baltimore, Maryland) crops out in a series of anticlines, mantled by mica schist, marble, and micaceous quartzite of the Glenarm Supergroup of Crowley (1976).[1] The older Precambrian rocks include metamorphosed intrusive rocks ranging from gabbro to granite and subordinate quartzose metasedimentary rocks (Wagner and Crawford, 1975), and are marked on the aeromagnetic map by conspicuous magnetic lows. The Glenarm metasedimentary rocks, chiefly mica schists of the Wissahickon Group of Crowley (1976)[2] and interlayered mafic and ultramafic rock, are marked by broad, open magnetic highs, having amplitudes as high as 1800γ.

The distribution of highs and lows on the magnetic map clearly outlines the areas where most of the Baltimore Gneiss crops out. The correspondence between the geologic map and the aeromagnetic pattern is especially clear for the West Chester anticline (Plates 2 and 3); even the small slice of Baltimore Gneiss protruding from the northwestern flank of this structure is clearly matched by an anomaly on the aeromagnetic map. A similar anomaly pattern is at Landenberg, Pennsylvania, and reconnaissance mapping guided by the aeromagnetic data has shown that here too the deep low is coextensive with Baltimore Gneiss exposed in a previously unmapped structure, the Mill Creek anticline, mantled by Glenarm metasedimentary rocks (Higgins and others, 1973).

The structure of the Baltimore Gneiss anticlines near Philadelphia—especially the southwestern ones—has proved controversial. Bailey and Mackin (1937) and Mackin (1950, 1962) pointed out that when viewed from the east-northeast, parallel to the plunge of the minor folds, the outcrop pattern of the Baltimore Gneiss suggests the presence of a faulted nappe, represented by the westernmost, hook-shaped (Woodville) anticline, and the southernmost (Avondale) anticline, folded over the west-southwest plunging West Chester anticline (Fig. 4). McKinstry (1961) has suggested an alternative interpretation, in which all the areas underlain by Baltimore Gneiss are considered as culminations on a complexly folded anticlinorium. The area underlain by Wissahickon schists be-

[1] Hereafter generally referred to as the "Glenarm".
[2] Hereafter generally referred to as the "Wissahickon".

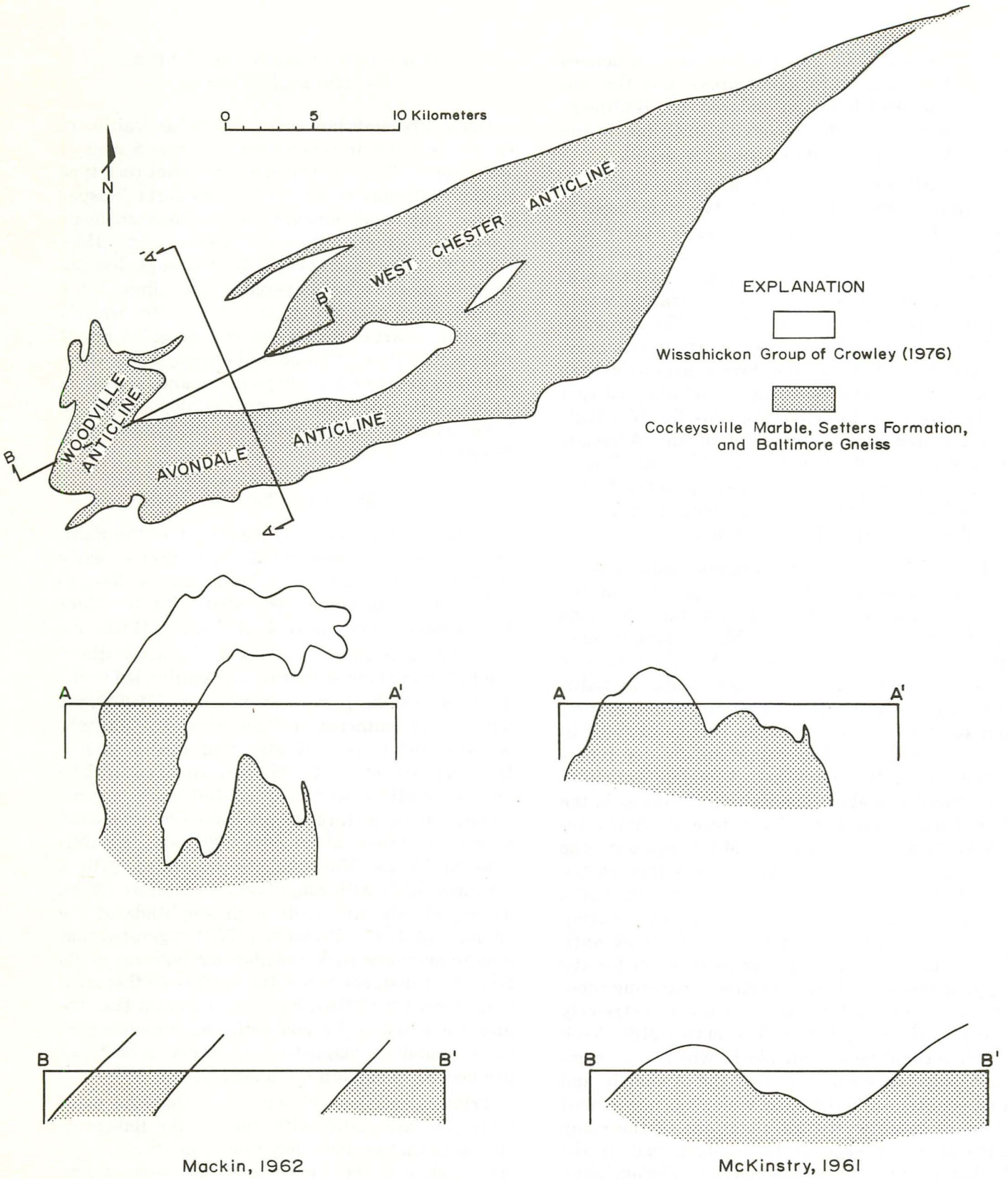


Figure 4. Geologic sketch map of gneiss anticlines west of Philadelphia, adapted from the Pennsylvania State geologic map (Gray and others, 1960). Differing interpretations of the structure by McKinstry (1961) and Mackin (1962) are shown. Nonmagnetic rocks (Cockeysville Marble, Setters Formation, and Baltimore Gneiss) are shown stippled; magnetic rocks (Wissahickon) are shown without overprint.

tween the Woodville and West Chester anticlines is visualized as a synformal depression in the center of the anticlinorium, floored by Baltimore Gneiss connecting the Woodville anticline to the West Chester anticline at a shallow depth.

As clearly brought out by Mackin (1962), an essential difference between the two hypotheses is the dip of the curved eastern contact of the Woodville anticline: Mackin's nappe interpretation would require this contact to plunge west-southwest, whereas McKinstry's interpretation would require it to plunge east-northeast. The aeromagnetic map (Plate 2) provides important data bearing on this question. The broad magnetic high associated with the Wissahickon schists just east of the Woodville anticline overlaps the Woodville anticline and the western end of the Avondale anticline, strongly suggesting that the Wissahickon schists plunge southwest, beneath the Woodville and Avondale anticlines, just as required by the nappe interpretation.

If the Woodville and Avondale anticlines do represent a refolded nappe, the position of the newly discovered Mill Creek anticline (Higgins and others, 1973) poses a problem. The aeromagnetic map shows no indication that the magnetic schists of the Wissahickon pass beneath the Baltimore Gneiss of the Mill Creek anticline, so presumably the Baltimore Gneiss here is exposed in a rooted antiform. The nappe apparently represented by the Woodville and Avondale anticlines lies structurally above the Baltimore Gneiss in the West Chester anticline. Therefore it should reappear structurally above the Mill Creek anticline as well; but there is no suggestion either on the published geologic maps or on the aeromagnetic map (Plate 2) of Baltimore Gneiss reappearing between the Avondale and the Mill Creek anticlines. There does not appear to be room for the nappe to pinch out in the syncline separating these two anticlines, unless the syncline is extremely deep and tightly pinched. The entire Mill Creek anticline could be a fault block, which was originally near the eastern end of the Avondale and West Chester anticlines, and which has now moved 20 km or more westward along one of the many faults that converge at the eastern end of the Baltimore Gneiss anticlines north of Philadelphia. Numerous other possibilities exist, of course, but resolution of this problem must await more detailed mapping in the Piedmont of Pennsylvania and Delaware.

Older Precambrian Rocks Near Baltimore, Maryland

Older Precambrian rocks known as Baltimore Gneiss crop out in seven areas north and west of Baltimore (Plate 3). The most abundant rock type in the Baltimore Gneiss is biotite-quartz-feldspar gneiss, but small amounts of biotite-hornblende gneiss and amphibolite are widespread; see Hopson (1964) and Olsen (1972) for thorough descriptions of these rocks. Layered amphibolites in the southwestern part of Baltimore City were formerly considered part of the Baltimore Gneiss, but recent radiometric dates (Tilton and others, 1970) show that they are considerably younger, and Southwick (1969), Higgins (1972), and Crowley (1976) all have assigned these rocks to the James Run Formation.

Magnetic Character

As noted by Bromery (1968) much of the Baltimore Gneiss is associated with deep, nearly featureless magnetic lows in the range 400γ to 500γ similar to those associated with the older Precambrian rocks near Philadelphia (Plate 2).

This magnetic expression is so characteristic of the Baltimore Gneiss that its recognition led to the discovery of the previously unknown Mill Creek anticline in southeastern Pennsylvania and northwestern Delaware (Higgins and others, 1973). However, exceptions to this pattern do occur. The gneiss anticlines southwest of Baltimore are associated with featureless anomalies of somewhat higher amplitude, about 600γ-800γ, and the anticline at Texas, Maryland, is associated with a magnetic high with amplitudes as high as 1300γ. To explain the unusually high amplitude of the anomaly at Texas, Bromery (1968) suggested that a more magnetic rock, possibly serpentinite, could be a short distance below the surface in this anticline. Crowley (1976), however, reported that the gneiss exposed at the surface in this anticline contains abundant magnetite; no other rock types are needed to explain the anomaly.

Trial calculations clearly show that the magnetic lows associated with most of the Baltimore Gneiss cannot be due solely to a susceptibility contrast between the gneiss and the surrounding Wissahickon schists; the average susceptibility of the Wissahickon schists (0.61×10^{-4} emu/cm^3; *see* Table 3) is relatively low so that even if the Baltimore Gneiss had zero susceptibility, the con-

trast in susceptibility between gneiss and schist can produce anomalies no larger than 10γ to 15γ, a full order of magnitude less than the observed anomaly. Consequently, we considered two possible explanations for the magnetic lows associated with most of the Baltimore Gneiss: (1) the gneiss could have reversed remanent magnetization, or (2) the gneiss could be structurally overturned, so that its normal remanent magnetization has been tectonically reversed.

In order to investigate these possibilities, we took 147 core samples from 25 sites in the Towson, Phoenix, and Woodstock anticlines (Table 1). These were drilled in bedrock using a water-cooled diamond-impregnated bit powered by a modified chain-saw motor.

Remanent magnetic moments were measured on a motor-driven "spinner" magnetometer (Doell and Cox, 1965) and all samples were magnetically "cleaned" in alternating current fields as high as 200 oersteds. Of the 147 samples, 122 had a significant direction of magnetization. Magnetic data for these samples are listed in Table 1.

The measured directions of remanent magnetization have a nearly random orientation, both within individual gneiss anticlines (Figs. 5, 6, 7) and for all three anticlines combined (Fig. 8). Consequently, the vector sum of the remanent directions approximates zero, probably accounting for the flat, featureless character of the magnetic anomalies associated with the Baltimore Gneiss. But this near-random orientation of magnetic poles implies that the remanent magnetization in the samples studied cannot be the cause of the observed lows. In addition, the measured values of induced magnetization (Table 1) confirm our earlier inference that the lows cannot reflect a susceptibility contrast between the gneiss and the surrounding schists; the induced magnetization data (Table 1) suggest that the susceptibility of the gneiss is on the order of 10^{-4} emu/cm^3, slightly *higher* than that of the nearby Wissahickon schists (*see* Table 3).

In short, the low magnetic anomaly associated with the northeastern gneiss anticlines at Baltimore cannot be explained in terms of the magnetic properties measured in our samples of Baltimore Gneiss. The anomalies must reflect the presence of a rock unit having reverse remanent magnetization—either in the gneiss itself or in the overlying Setters Formation or Cockeysville Marble—which was not detected in our sampling program. The failure of our sampling to find this unit suggests that it is thin and can be detected only by a much more extensive sampling and drilling program. But despite our uncertainty as to the origin of these anomalies, the strong association of the magnetic lows with the mapped extent of the gneiss anticlines (Plates 1 and 2) and especially our success in using them to discover a previously unmapped gneiss anticline (Higgins and others, 1973) indicate that the anomaly patterns are useful guides to the form and extent of the gneiss anticlines.

Geologic Interpretation

Since the pioneering work of Broedel (1937), no systematic studies of the structure of the Baltimore Gneiss anticlines have been made. The swirling map pattern of the Precambrian rocks suggests a fascinating history of repeated folding, but thus far none of the details have been worked out. Fortunately however, the aeromagnetic data provide some help in deciphering the large-scale form of these anticlines. The Wissahickon which lies stratigraphically above the Baltimore Gneiss, is associated with magnetic anomalies of moderate amplitude having a distinctive, irregular pattern which contrasts markedly with the magnetic expression of the Baltimore Gneiss (Plate 2). As discussed by Bromery (1968, p. 76), anomalies associated with the Wissahickon extend across the contacts into the Baltimore Gneiss at several places, suggesting that the contacts are overturned, the magnetic Wissahickon schists underlying Baltimore Gneiss and Cockeysville Marble. These relations are best shown along the northern flanks of the Towson and Chattolanee anticlines, and on the southeastern side of the Woodstock anticline. The anomaly produced by the Wissahickon extends entirely across the Woodstock anticline, to its southwest plunging nose, suggesting that this structure is a southwest-plunging refolded sheet, rather than a simple dome.

Another important anomaly recognized by Bromery (1968) is in the central part of the Phoenix anticline, the northernmost of the anticlines near Baltimore. The northwestern part of this anticline encloses an elongate patch of Glenarm metasedimentary rocks, conventionally interpreted as a synformal "dimple" on the crest of the anticline (Knopf and Jonas, 1929, p. 186). Mica schists of the Wissahickon, which crop out in the center of this patch, are marked by a prominent twin-peaked magnetic high (Plate 2). A high centered over the Cockeysville Marble just

TABLE 1.—*Magnetic data for samples from the Baltimore Gneiss and overlying Setters Formation*
(emu, electromagnetic units)

Sample No.	Induced magnetization ($emu/cm^3 \times 10^{-5}$)	Remanent magnetization ($emu/cm^3 \times 10^{-5}$)	Q ratio Remanent magnetization / Induced magnetization	Metamorphic rock
BG1-1	52.40	7.24	.14	Baltimore Gneiss
BG1-2	16.10	6.44	.40	do
BG1-3	7.10	2.61	.37	do
BG1-4	20.50	10.47	.51	do
BG1-5	1.70	.43	.25	do
BG1-6	1.70	< .40	< .23	do
BG1-7	3.30	1.71	.52	do
BG1-8	6.08	2.47	.61	do
BG2-1	2.80	2.47	.88	do
BG2-2	2.60	.43	.16	do
BG2-3	4.00	252.00	63.00	do
BG2-4	1.70	114.70	67.40	do
BG2-5	3.40	23.4	6.87	do
BG3-1	1.70	< .40	< .23	do
BG3-2	1.70	< .40	< .23	do
BG3-3	1.70	< .40	< .23	do
BG3-4	1.70	< .40	< .23	do
BG3-5	1.70	< .40	< .23	do
BG4-1	1.70	< .40	< .23	do
BG4-2	8.40	46.90	5.59	do
BG4-3	3.80	24.30	6.40	do
BG4-4	8.80	5.20	.59	do
BG4-5	2.30	1.85	.81	do
BG5-1	1.70	.57	.34	Baltimore Gneiss
BG5-1	1.70	.43	.25	do
BG5-3	1.70	< .40	< .23	do
BG5-4	1.70	< .40	< .23	do
BG5-5	1.70	< .40	< .23	do
BG6-1	1.70	< .40	< .23	do
BG6-2	1.70	< .40	< .23	do
BG6-3	1.70	< .40	< .23	do
BG7-1	1.70	< .40	< .23	Setters Formation
BG7-2	1.70	< .40	< .23	do
BG7-3	1.70	< .40	< .23	do
BG7-4	1.70	< .40	< .23	do
BG8-1	10.50	17.70	1.70	Setters Formation
BG8-2	6.40	2.70	.41	do
BG8-3	10.50	2420.00	231.00	do
BG8-4	8.30	48.30	5.80	do
BG8-5	6.60	373.50	56.60	do
BG9-1	2.80	1.50	.54	Baltimore Gneiss
BG9-1	8.60	.96	.11	do
BG9-3	.05	.90	.18	do
BG9-4	.03	.73	.22	do
BG9-5	.02	164.00	7130.00	do
BG10-1	1.90	.43	.23	do
BG10-2	4.60	.96	.21	do
BG10-3	3.90	1.13	.29	do
BG10-4	2.30	1.29	.56	do

TABLE 1.—*Magnetic data for samples from the Baltimore Gneiss and overlying Setters Formation*—(continued)
(emu, electromagnetic units)

Sample No.	Induced magnetization (emu/cm³ x 10⁻⁵)	Remanent magnetization (emu/cm³ x 10⁻⁵)	Q ratio Remanent magnetization / Induced magnetization	Metamorphic rock
BG11-1	6.60	59.30	9.00	Baltimore Gneiss
BG11-2	4.70	1.21	.26	do
BG11-3	4.30	1.10	.24	do
BG11-4	6.80	4.62	.68	do
BG11-5	2.30	5.15	22.28	do
BG11-6	1.90	4.40	2.31	do
BG12-1	1.70	< .40	< .23	do
BG12-2	5.10	234.60	46.00	do
BG12-3	1.90	.91	.48	do
BG12-4	7.40	462.00	62.40	do
BG12-5	16.50	1002.00	60.80	do
BG13-1	2.80	.66	.24	do
BG13-2	2.30	.50	.22	do
BG13-3	1.90	< .40	< .21	do
BG13-4	1.90	.43	.23	do
BG13-5	16.50	1002.00	60.80	do
BG14-1	8.40	10.40	1.20	do
BG14-2	8.00	10.90	1.40	do
BG14-3	6.80	8.20	1.20	do
BG14-4	2.30	.73	.32	do
BG14-5	3.40	493.00	145.00	do
BG15-1	7.90	16.40	2.10	do
BG15-2	4.70	5.70	1.21	do
BG15-3	2.10	6.01	2.86	do
BG15-4	2.10	.71	.34	do
BG15-5	2.10	5.64	2.68	do
BG16-1	1.90	39.50	20.80	Setters Formation
BG16-2	3.40	65.00	19.10	do
BG16-3	1.90	32.90	17.30	do
BG17-1	5.20	14.40	2.80	Baltimore Gneiss
BG17-2	5.20	2.69	.52	do
BG17-3	5.20	.60	.12	do
BG17-4	3.90	.67	.17	do
BG17-5	1.70	.43	.25	do
BG18-1	2.10	.90	.43	do
BG18-2	1.80	.43	.24	do
BG18-3	1.70	.43	.25	do
BG18-4	1.70	.43	.25	do
BG19-1	4.30	3.79	.88	do
BG19-2	17.40	25.30	1.40	do
BG19-3	9.70	3.25	.34	do
BG19-4	5.10	.75	.15	do
BG19-5	3.80	2.11	.56	do
BG20-1	1.70	.43	.25	do
BG20-2	1.70	< .40	< .23	do
BG20-3	1.70	.43	.25	do
BG20-4	1.70	< .40	<.23	do
BG20-5	1.70	.43	.25	do
BG21-1	5.20	< .40	< .08	do
BG21-2	5.20	< .40	< .08	do
BG21-3	5.20	< .40	< .08	do
BG21-4	5.20	< .40	< .08	do
BG21-5	5.20	< .40	< .08	do

TABLE 1.—*Magnetic data for samples from the Baltimore Gneiss and overlying Setters Formation*—(continued)
(emu, electromagnetic units)

Sample No.	Induced magnetization ($emu/cm^3 \times 10^{-5}$)	Remanent magnetization ($emu/cm^3 \times 10^{-5}$)	Q ratio Remanent magnetization / Induced magnetization	Metamorphic rock
BG22-1	4.10	9.06	2.30	Baltimore Gneiss
BG22-2	2.30	.37	.16	do
BG22-3	1.90	.71	.37	do
BG22-4	5.50	2.19	.40	do
BG22-5	6.60	4.95	.75	do
BG23-1	5.50	29.10	5.30	do
BG23-2	5.00	2.76	.55	do
BG23-3	4.30	1.78	.41	do
BG23-4	3.40	.67	.20	do
BG23-5	3.40	.88	.26	do
BG24-1	48.30	10.80	.22	do
BG24-2	3.50	2.31	.66	do
BG24-3	3.30	.65	.20	do
BG24-4	29.90	18.30	.61	do
BG24-5	18.80	6.81	.36	do
BG24-6	4.60	1.81	.39	do
BG24-7	50.90	53.80	1.10	do
BG25-1	3.40	< .40	< .12	do
BG25-2	3.00	.70	.23	do
BG25-3	4.60	2.08	.45	do

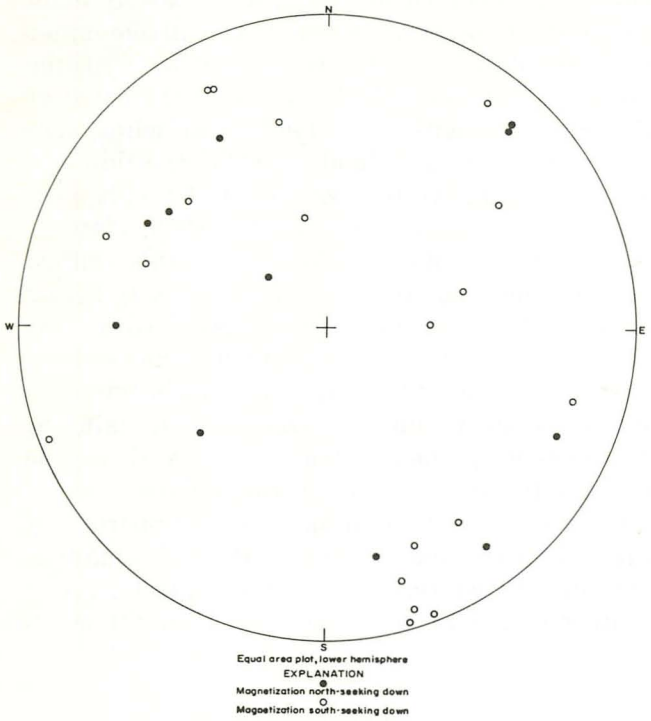


Figure 5. Directions of remanent magnetization for gneiss samples from the Towson Anticline, Baltimore County, Maryland.

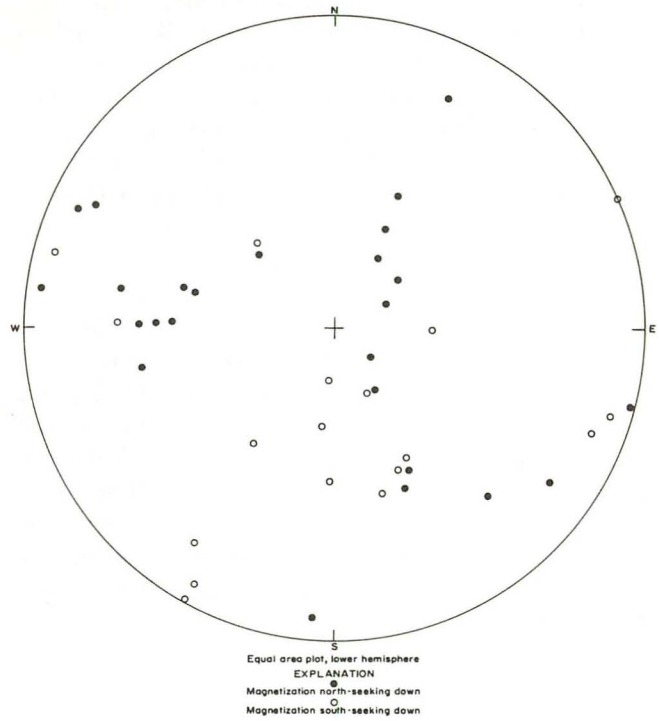


Figure 6. Directions of remanent magnetization for gneiss samples from the Woodstock Anticline, Howard County, Maryland.

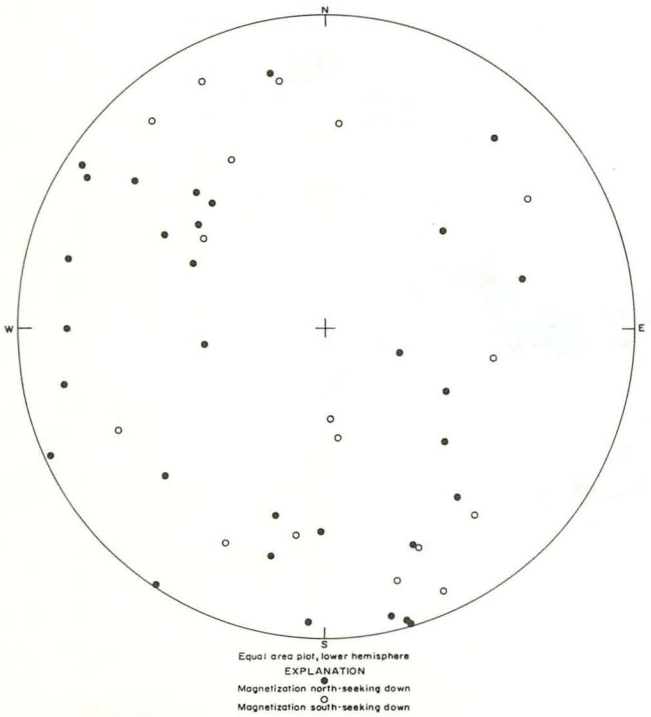


Figure 7. Directions of remanent magnetization for gneiss and quartzite samples from the Phoenix Anticline, Baltimore County, Maryland.

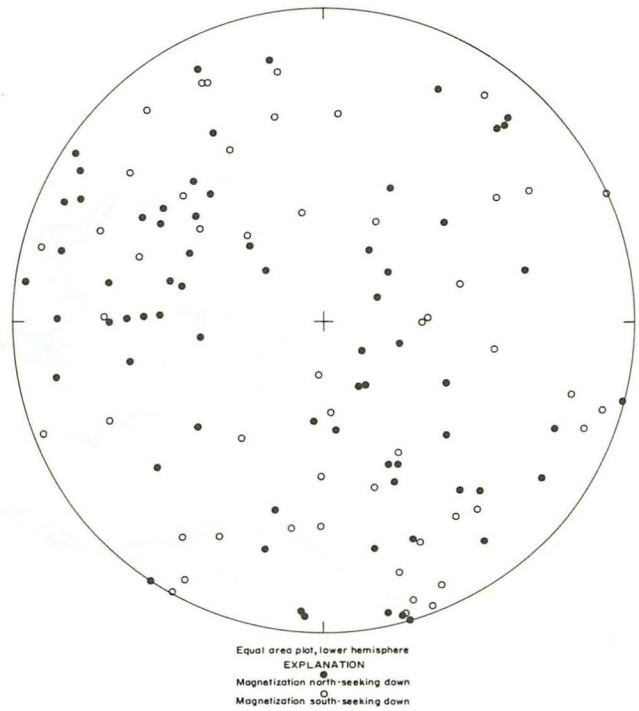


Figure 8. Summary of directions of remanent magnetization for gneiss and quartzite samples from the Towson, Phoenix, and Woodstock anticlines, Baltimore and Howard Counties, Maryland.

3 km to the east closely resembles the anomaly over the Wissahickon. Bromery (1968, p. 73) suggests that this anomaly could reflect Wissahickon schists structurally beneath the Cockeysville; if so, the Cockeysville here, and presumably the Wissahickon schists just to the west, must be exposed on the crest of an antiform, bringing up rocks structurally beneath the Baltimore Gneiss. As noted by Bromery, the gradients in this anomaly decrease substantially northeastward, suggesting that the rocks producing the anomaly plunge northeastward. In view of this pattern, it seems likely that the gneiss in the Phoenix anticline is the tongue of a recumbent fold, having Glenarm metasedimentary rocks both above and below the gneiss, which has been refolded about a northeast-trending set of axes. Figure 9 shows a possible cross section of the inferred structure, drawn down the plunge of minor folds in the northeast part of the anticlinorium centered near Baltimore.

Final evaluation of this intriguing possibility must await detailed mapping, and possibly drilling, of the Phoenix anticline, but available minor structural data are perfectly compatible with the interpretation shown in Figure 9. If the patch of Glenarm rocks within the Phoenix anticline were synformal, fold axes should parallel the thin belt of gneiss at the western end of the Phoenix anticline, but they do not. Choquette (1960, Plate 3) showed about 2 dozen fold axes in this critical area, all plunging west-southwest, directly across the belt of gneiss, precisely as required by the interpretation of Figure 9. Furthermore, Choquette's map shows that the fold axes in the small belt of Cockeysville apparently underlain by Wissahickon plunge northeast, parallel to the plunge inferred from the magnetic data.

The nappe interpretation is also supported by lithologic variation in the Setters Formation; Crowley (1976) reported that the garnet schist member of the Setters occurs on both the north

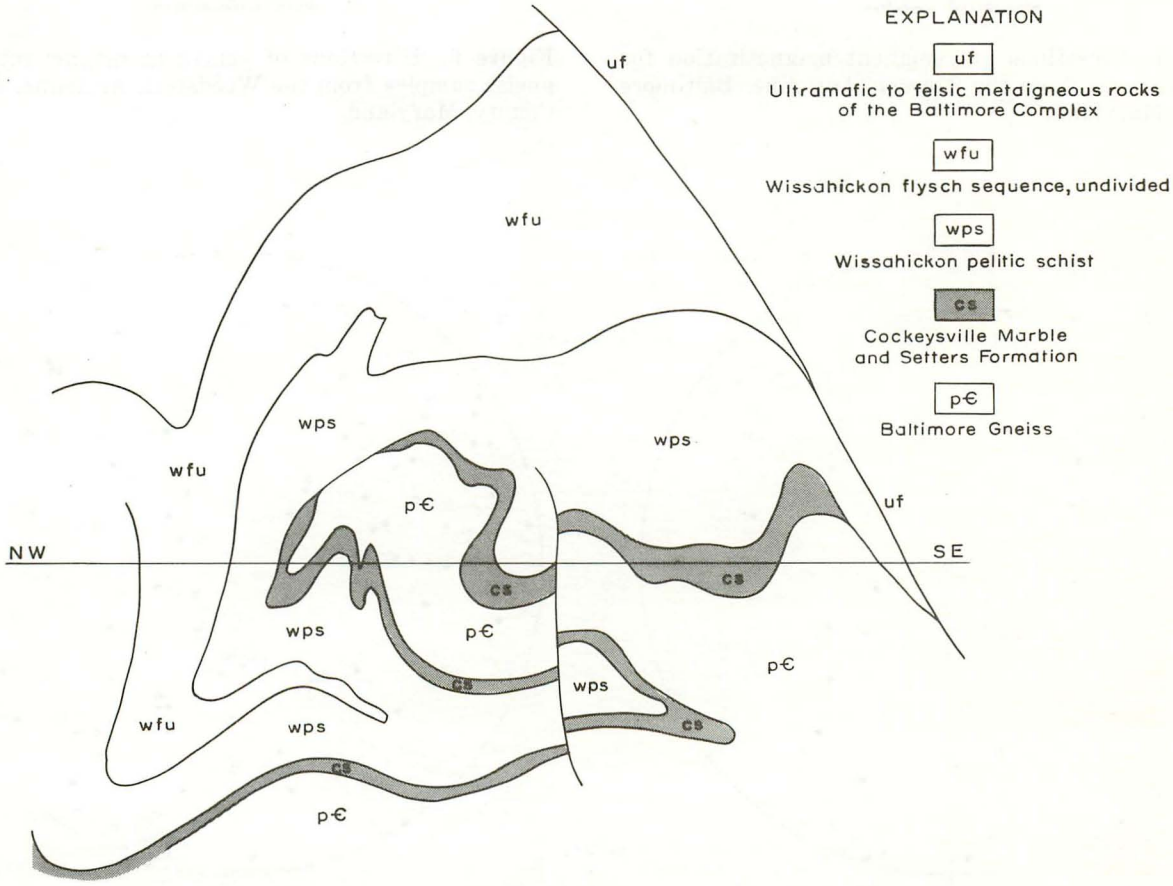


Figure 9. Interpretative cross section of gneiss anticlines near Baltimore, constructed by viewing the geologic map down the plunge of second generation minor folds in the northeastern part of the anticlinorium at Baltimore. Contacts adapted from the State geologic map of Maryland (Cleaves and others, 1968).

and south flanks of the Phoenix anticline, but does not occur in the inlier of Glenarm metasedimentary rocks. Consequently, the synformal interpretation of the infolded belt of Glenarm rocks would require the startling coincidence that the fold crest along the western end of the Phoenix dome must exactly coincide with a facies change in the Setters Formation. Given the sinuous form of the fold crest, this relation seems highly unlikely. On the other hand, if the nappe interpretation is correct, the quartzites exposed inside the Phoenix anticline may have been deposited many kilometers away from the Setters schists exposed outside the anticline, and a marked lithologic difference between the two is easily understood.

During our traverses of the Phoenix anticline, we found an outcrop area of potassic quartzite, which resembles quartzite of the Setters Formation, within the Phoenix anticline near Glencoe, Md. Primary layering in the Baltimore Gneiss dips away from this quartzite suggesting that the Setters is here exposed in another window through the refolded nappe.

Conclusions

In each of the three areas we have examined, the aeromagnetic data have confirmed or suggested the presence of major refolded nappe structures. In this region of low relief and poor exposures, the three-dimensional insight that aeromagnetic data can provide in interpreting the geometry of these large-scale structures has proved invaluable. In order to document the large-scale structures more completely, more detailed magnetic surveys, using ground and/or truck-mounted magnetometers, coupled with detailed mapping of all available minor structures should prove extremely helpful. Areas which appear particularly well suited to this sort of study are the Phoenix and Woodstock anticlines near Baltimore, the Woodville anticline west of Philadelphia, and the Green Pond syncline in the Reading Prong.

THE GLENARM AND RELATED ROCKS

The Glenarm Supergroup of Crowley (1976), a thick section of predominantly clastic metasedimentary rocks (Table 2), crops out in a broad northeast-trending synclinorium bounded on the west by the chain of anticlinoria stretching from South Mountain in Maryland to Mine Ridge in Pennsylvania, and on the east by the line of gneiss anticlines culminating at Baltimore and Philadelphia (Plate 3). Interpreting the age and correlation of these rocks has been one of the principal problems of Piedmont geology. The evolution of thought on this question is long and complex (*see* review in Higgins, 1972), but most authors have favored one or the other of two interpretations: 1) that the Glenarm is mainly a eugeosynclinal facies more or less equivalent in age to the upper Precambrian-lower Paleozic miogeosynclinal section exposed west of South Mountain and Mine Ridge; or 2) that the Glenarm is an upper Precambrian sequence, possibly related to the Ocoee Supergroup or the Lynchburg Formation of the southern Appalachians but, in any case, older than the Paleozoic section to the northwest; most who have favored this latter interpretation have mapped the contact between the Piedmont section and the Paleozoic rocks as a major thrust fault, the Martic Thrust.

Distinguishing between these two possibilities is obviously pivotal to any interpretation of the evolution of the central Appalachians. One of the principal problems has been the difficulty of mapping the schists and phyllites of the Glenarm Supergroup in the central part of the synclinorium between South Mountain and Baltimore. Numerous difficulties are involved in mapping this area: the rocks are unfossiliferous, poorly exposed, complexly deformed, and variably metamorphosed. But the most frustrating problem has been the lithologically monotonous character of the rocks and the virtual absence of distinctive marker units. Fortunately, however, the aeromagnetic map shows many anomalies in this area that appear to outline recognizable geologic structures, suggesting the presence of units that are at least magnetically distinctive (Plate 2). Therefore, we have tried to identify these units, and use them to work out a provisional picture of Piedmont structure and stratigraphy.

In the following sections, we first discuss the magnetic properties of the rocks in the Glenarm and attempt to define a set of magnetically recognizable stratigraphic units. We then use these units to interpret the structure of the Piedmont synclinorium. Finally, we attempt to work out the relation between the stratigraphic units in the Glenarm and those in the South Mountain section.

TABLE 2

Stratigraphic Nomenclature in the Glenarm Supergroup

This Report GLENARM SUPERGROUP OF CROWLEY (1976)	*Crowley, 1976* GLENARM SUPERGROUP	Higgins, 1972 GLENARM SERIES
Peach Bottom Slate	(not exposed in area mapped)	Cardiff Metaconglomerate[1]
Cardiff Metaconglomerate[1]		
James Run Formation[2]	(recognized, but not included in Glenarm Supergroup)	James Run Formation
Wissahickon Group of Crowley (1976)	Wissahickon Group	Wissahickon Formation
Diamictite	Sykesville Formation	Diamictite facies
	Gneiss member	
	Schist member	
	(not recognized; included with Oella Formation)	Quartzite facies
Metagraywacke	Oella Formation	
	Sweathouse amphibolite member	Metagraywacke facies
	Piney Run Formation	
Quartzose schist	Pleasant Grove Schist	
Pelitic schist	Loch Raven Schist	Pelitic schist facies
	Hydes marble member	
	Rush Brook member	
	Prettyboy Schist	
Cockeysville Marble	Cockeysville Marble	Cockeysville Marble
	Massive metadolostone member	
	Massive metalimestone member	
Undifferentiated	Layered metadolostone member	Undifferentiated
	Layered marble member	
	Phlogopitic metalimestone member	
Setters Formation	Setters Formation	Setters Formation
Garnet schist member	Garnet schist member	
	Quartzite member	
Undifferentiated	Schist lens	Undifferentiated
	Conglomerate lens	
	Gneiss member	
	Quartzite lens	

[1] The stratigraphic position of the Cardiff Metaconglomerate and the Peach Bottom Slate is uncertain; current U. S. Geological Survey usage follows that of Higgins (1972), who interpreted the Cardiff as overlying the Peach Bottom, and underlying the Wissahickon.

[2] Stratigraphic relations between the James Run Formation and the Wissahickon are complex. The James Run appears to be younger than some Wissahickon schists in northeastern Maryland and is probably equivalent in age to other rocks in the Wissahickon farther west (Higgins, 1972); it is here interpreted as older than the uppermost part of the Wissahickon.

Magnetic Properties and Distribution of Rock Types

Setters Formation

Crowley (1976) has divided the Setters Formation into three lithologic members which are characterized by quartzite, gneiss, and garnet schist, respectively (Table 2); his garnet schist member corresponds closely to the kyanite-, staurolite-, and garnet-bearing schist of the Setters described by Fisher (1971). As mapped by Crowley (1976) in Baltimore County, Md., the quartzite and gneiss members greatly predominate; the garnet schist member is restricted to two lenses in the Phoenix anticline and a thin layer wrapping around the Texas anticline (Plate 3).

On the aeromagnetic map (Plate 2), the quartzite and gneiss of the Setters are everywhere associated with flat, low magnetic anomalies indistinguishable from those produced by the Baltimore Gneiss, and clearly reflecting the low magnetic susceptibility of the rocks; measured magnetic susceptibilities in the quartzite and gneiss measured at two outcrops are less than 0.10×10^{-4} emu/cm^3 (Table 3). However, both lenses of the garnet schist member in the Phoenix anticline are associated with pronounced magnetic highs having amplitudes as high as 1000γ, reflecting the higher magnetic susceptibilities of these schists; measured susceptibilities range from 0.10×10^{-4} to 0.50×10^{-4} emu/cm^3 and average 0.30×10^{-4} emu/cm^3 (Table 3). Bromery (1968) recognized these anomalies, and their relation to the garnet-bearing schists, but interpreted the schists as infolded belts of the overlying Wissahickon Group, which they resemble lithologically. However, their consistent stratigraphic position in the Setters Formation, just beneath the Cockeysville (Fisher, 1971; Crowley, 1976) clearly suggests that these schists are a part of the Setters Formation.

The anomaly associated with the lens of garnet-mica schist on the south rim of the Phoenix anticline extends southward in an arc curving beneath the Cockeysville; kyanite-garnet-mica schist was encountered 116 m beneath the Cockeysville in drilling done by the Harry T. Campbell Co. at this locality (Fisher, 1971), and the anomaly clearly reflects the subsurface extent of the Setters schists. As noted by Bromery (1968, p. 73), the southward decrease in gradient shown by this anomaly reflects the southward increase in depth to garnetiferous schists. The belt of garnet-mica schists surrounding the Texas anticline appears to have no magnetic expression probably because any anomaly produced by the Setters schists is too small to be distinguished from the larger anomaly produced by the highly magnetic gneisses in the Texas anticline.

On the basis of these results, the aeromagnetic map appears to provide a useful guide to the distribution of the garnet schist member within the Setters Formation. No marked magnetic anomalies are associated with the Setters Formation in Howard County, Maryland, or in the anticlines west of Philadelphia; therefore, the garnet schist member is probably restricted to the northwesternmost part of the Setters, as suggested by Fisher (1971).

Cockeysville Marble

Crowley (1976) has divided the Cockeysville Marble into five lithologic members of uncertain stratigraphic significance. The aeromagnetic map (Plate 1) shows no magnetic anomalies attributable to any of the rock types in the Cockeysville; all are characterized by low, flat magnetic anomalies indistinguishable from those produced by most parts of the Baltimore Gneiss or by quartzite and gneiss of the Setters Formation. Measurements of the magnetic susceptibility at one locality are less than 0.10×10^{-4} emu/cm^3 (Table 3).

Wissahickon Group of Crowley (1976)

Crowley (1976) elevated the former Wissahickon Formation to group status, and subdivided it into six formations and five members, on the basis of his mapping in Baltimore County, Maryland. We recognize the desirability of these subdivisions for detailed mapping purposes, but in this reconnaissance study we have been unable to distinguish among several of Crowley's (1976) formations. For our purposes, we find the earlier lithologic nomenclature of Southwick and Fisher (1967), Higgins and Fisher (1971), and Higgins (1972) more useful. Accordingly, we will use a system of nomenclature similar to that defined by Higgins (1972), except that we will follow Crowley (1976) in referring to the Wissahickon as a group. The relation between our nomenclature and that of Higgins (1972) and Crowley (1976) is given in Table 2.

The diamictite facies of the Wissahickon Group (Sykesville Formation of Crowley, 1976) is associated with irregular magnetic highs having amplitudes as high as 2000γ. The form of these anomalies clearly parallels regional structural trends and outlines many details of local struc-

TABLE 3

Magnetic Susceptibilities of Piedmont Rocks

Rock Unit	Average Susceptibility emu/cm³ [1]	Standard Deviation	Outcrops[2] Measured
Wissahickon Group of Crowley (1976)			
Flysch sequence			
Diamictite	30.7×10^{-4}	18.0×10^{-4}	15
Metagraywacke	27.0×10^{-4}	22.2×10^{-4}	6
Quartz schist	24.3×10^{-4}	8.1×10^{-4}	3
(entire flysch sequence)	28.99×10^{-4}	17.65×10^{-4}	24
Pelitic Sequence			
Garnet-mica schist	0.56×10^{-4}	0.52×10^{-4}	8
Albite-chlorite phyllite	0.85×10^{-4}	0.14×10^{-4}	3
Ijamsville phyllite	0.56×10^{-4}	0.26×10^{-4}	8
(entire pelitic sequence)	0.61×10^{-4}	0.38×10^{-4}	19
Sams Creek Formation	2.89×10^{-4}	3.04×10^{-4}	15
Cockeysville Marble	0.10×10^{-4}	——	1
Setters Formation			
Quartzite and gneiss	0.10×10^{-4}	——	2
Garnet-mica schist	0.30×10^{-4}	——	2

[1] emu are electromagnetic units.

[2] At most outcrops, approximately ten measurements were made by means of a Bison field bridge, then averaged; at a few outcrops only two or three measurements could be obtained.

tures. *In situ* measurements of magnetic susceptibility give values ranging from 5.8 to 75.0×10^{-4} emu/cm³, and average 30.7×10^{-4} emu/cm³ (Table 3), reflecting the high magnetite content of these rocks (*see* modal analyses in Hopson, 1964). Local variations in the intensity of the anomalies associated with the diamictites and the large range in measured susceptibilities suggest substantial local variations in the magnetite content.

The metagraywacke facies of the Wissahickon (Crowley's (1976) Piney Run Schist and part of his Oella Formation) is associated with magnetic anomalies having a pattern and intensity much like those associated with the diamictite facies, and they have very nearly the same magnetic susceptibility (Table 3). These lithologies appear to be indistinguishable magnetically.

The pelitic schist facies of Higgins and Fisher (1971) included a variety of rock types, which we will treat separately. On the southeast it included coarse garnet-mica schists (Loch Raven Schist of Crowley, 1976) which lie stratigraphically between the Cockeysville Marble and the diamictite facies of the Wissahickon (Plate 3). These rocks are associated with magnetic anomalies having an irregular, "birds-eye maple" pattern, and an amplitude of 800 to 1200γ, intermediate between the anomaly amplitudes characteristic of the Cockeysville and those associated with the diamictite facies (Plate 2). *In situ* magnetic susceptibilities range from 0.10×10^{-4} to 1.50×10^{-4} emu/cm³, also intermediate between those of the Cockeysville and those of the diamictite facies (Table 3).

Petrographically similar schists in the same stratigraphic positions near the West Chester anticline in Pennsylvania have quite a different magnetic expression (Plate 1); they are characterized by smooth, regular anomalies of much higher amplitude (as high as 2000γ). To judge from descriptions given by Bascom and Stose (1932) and Hopson (1964), these schists closely resemble those at the base of the Wissahickon near Baltimore, and we believe that the two are stratigraphically correlative. The different magnetic expression could reflect a higher magnetite content in the Pennsylvania schists, possibly due to a higher ferric iron content. Alternatively, Bascom and Stose (1932) showed many small (and some large) pods of serpentine and gabbro interlayered with the Pennsylvania schists, and the magnetic anomalies could reflect the presence of these bodies, rather than higher susceptibilities of the schists themselves.

Higgins and Fisher (1971) showed the area underlain by Wissahickon northwest of the diamictite-metagraywacke belt as undifferentiated pelitic schist. However, the map pattern of mag-

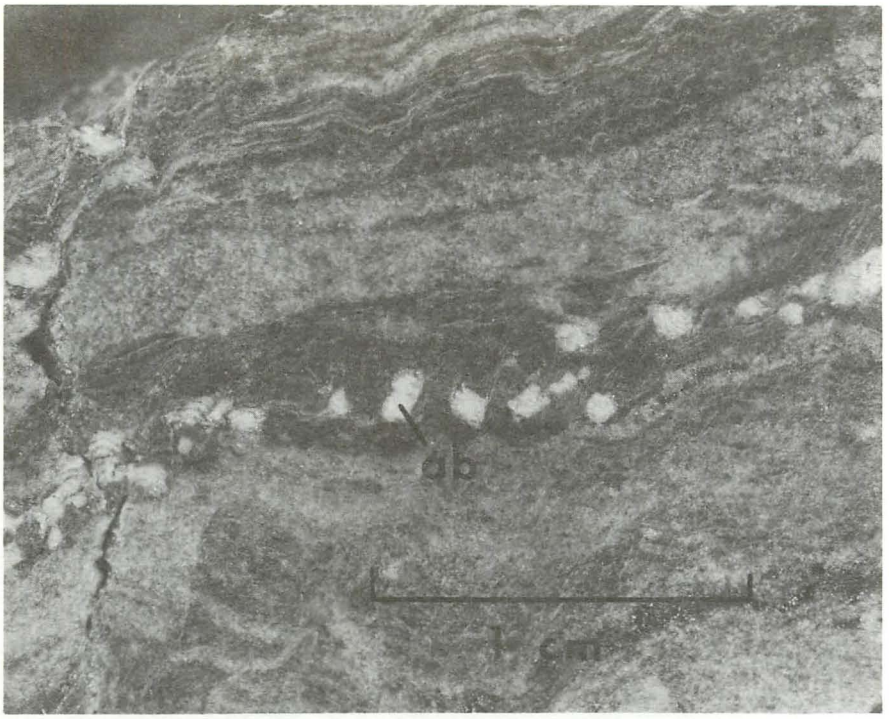


Figure 10. Photograph of polished slab of albite-chlorite phyllite in Wissahickon. Note euhedral albite porphyroblasts (ab) in matrix composed of alternating layers of chlorite-rich phyllite (dark layers) and muscovite-rich phyllite (light gray layers). Sample from along Western Maryland Railroad tracks, 3.6 km southwest of Westminster, Carroll County, Maryland.

netic anomalies in this area (Plate 2) clearly reveals units of differing magnetic susceptibilities, and suggests the presence of refolded folds.

Rapid reconnaissance mapping guided by the aeromagnetic maps suggests that this area is underlain principally by two rock types, each characterized by a relatively uniform magnetic expression. Most of the areas associated with lows on the aeromagnetic map are underlain by porphyroblastic albite-chlorite phyllites (Figure 10). Southwick (1969, p. 39) and Stose and Jonas (1939) have described these rocks from within the magnetic low near Norrisville, Maryland. Crowley (1976) reports this rock type (his Prettyboy Schist) from the area of the northeast-trending magnetic low just east of Hampstead, Maryland; and Fisher (1978) has mapped the same lithology in the featureless magnetic low east of Westminster, Maryland. Each of these areas is characterized by a relatively featureless magnetic pattern having amplitudes in the range 600γ to 1000γ and measured susceptibilities in the range 0.73×10^{-4} to 1.01×10^{-4} emu/cm^3 (Table 3).

In the northwesternmost part of the Piedmont in Maryland and Pennsylvania phyllite traditionally assigned to the Ijamsville Phyllite and "Marburg Schist" (of local usage) has the same magnetic expression and the same range of magnetic susceptibilities as the albite-chlorite schists of the Wissahickon (Plate 2, Table 3). Fisher (1978) recently showed that the Ijamsville and "Marburg" were equivalent units and proposed dropping the name "Marburg"; we shall do so here. Furthermore, Fisher showed that the Ijamsville and the albite-chlorite phyllites of the Wissahickon are stratigraphically equivalent, and differ only in metamorphic grade. Because we wish to examine stratigraphic relations in the Piedmont, rather than variations in metamorphic grade, we will not differentiate between the albite-chlorite phyllites of the Wissahickon Group and the Ijamsville (or "Marburg") Phyllite. In Plate 3, all of these units are mapped as pelitic schists of the Wissahickon.

Many of the areas characterized by magnetic highs in the northwestern Piedmont are under-

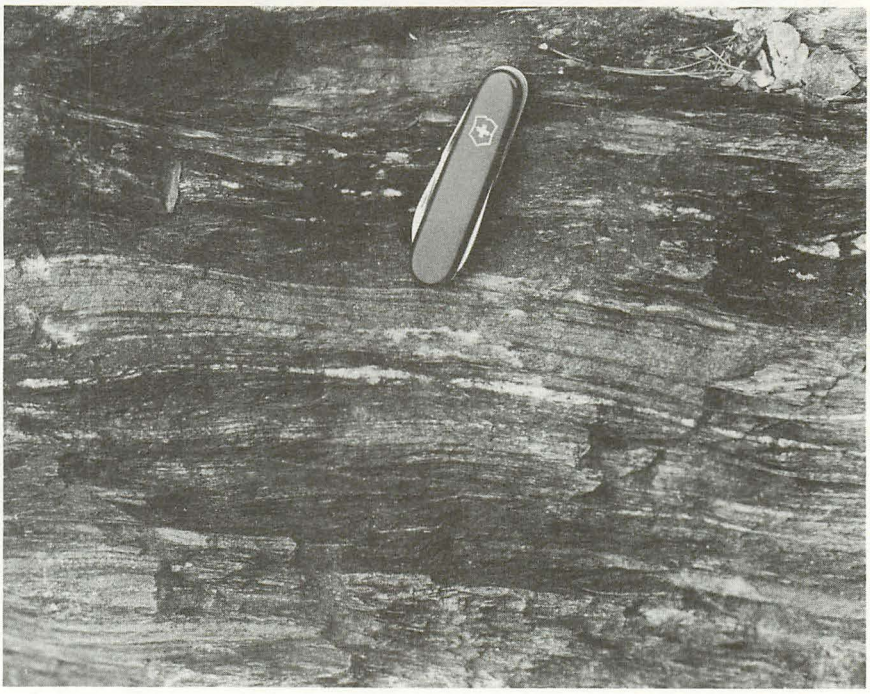


Figure 11. Laminated mica-quartz schist of Wissahickon. Knife handle is 8.3 cm long. Outcrop at intersection of Masemore Road and Gunpowder Falls, 3.0 km northwest of Hereford, Baltimore County, Maryland.

lain by a distinctive quartzose schist, lacking prominent albite porphyroblasts and containing delicately laminated quartzose beds (Figure 11). Crowley (1976) has mapped some of these rocks as "Pleasant Grove Schist" in the area characterized by strong, linear magnetic highs in northwestern Baltimore County. In most places, the contact between his Pleasant Grove Schist and his Prettyboy albite-chlorite schist corresponds well with the break on the magnetic map between anomalies greater than 1000γ and those less than 1000γ (Plate 2). Southwick and Owens (1969) mapped a contact between schists containing prominent albite porphyroblasts on the northwest, and schists that are finer grained, and more equigranular on the southeast; this line coincides almost exactly with the break on the magnetic map between anomalies with amplitudes greater than 1000γ and those having lesser amplitudes. Accordingly, we conclude that Southwick's fine-grained, equigranular schists are part of the same lithologic unit as Crowley's Pleasant Grove Schist. Rocks characterized by similar magnetic expression and lithology have long been mapped as Peters Creek Quartzite in Pennsylvania; most of these rocks are probably equivalent to Crowley's Pleasant Grove. Hopson (1964, p. 88) reported lithologically similar rocks in the Wissahickon schists about 20 km northwest of Washington, D.C., near the contact with the albite-chlorite phyllite of the Wissahickon. Once again the break in magnetic pattern follows this contact almost perfectly (Plate 2).

Because of the consistent relationship between magnetic expression and lithology in each of the areas we have examined, we suggest that the schists of the northwestern part of the Piedmont can be subdivided into two magnetically distinguishable lithologic types: (1) fine-grained, weakly magnetic, prophyroblastic albite-chlorite phyllites and schists of the Wissahickon, and equivalent paragonite-chlorite phyllites, formerly assigned to the Ijamsville Phyllite; and (2) quartzose schist of the Wissahickon Group which commonly lacks albite porphyroblasts, contains thin, delicately laminated beds of micaceous quartzite, and is strongly magnetic. Measured magnetic susceptibilities for both rock types are given in Table 3.

In Plate 3 we have attempted to show the extent

Figure 12. Metagraywacke interbedded with mica-quartz schist in the Wissahickon. Note that metagraywacke is graded and has massive base, locally balled up into load casts, and laminated top. Top of beds is toward top of picture. Hammer handle is 40 cm long. Outcrop along the Potomac River, at Bear Island, Montgomery County, Maryland.

of these two units in the western part of the Piedmont, on the basis of published maps, our own reconnaissance, and the distribution of high-amplitude and low-amplitude anomalies on the aeromagnetic maps (Plate 2). In drawing this map, we have included the "Octoraro Schist" (local usage) of Pennsylvania, and the Urbana Phyllite of Maryland with the non-magnetic Wissahickon. Both units resemble the non-magnetic Wissahickon phyllites in both magnetic expression and lithology, but we have not studied either unit in detail.

The contact between the Wissahickon Group and the fossiliferous lower Paleozoic rocks flanking the Piedmont on the northwest has long been controversial. Traditionally known as the Martic Line, this contact has been interpreted as a regional thrust fault, a local thrust fault, and a normal stratigraphic contact (*see* reviews in Wise, 1970, and Higgins, 1972, for a full account of the history of this controversy). Part of the problem has been the difficulty of mapping the contact in the area between Hanover, Pennsylvania and the Mine Ridge anticline, where the Wissahickon is in contact with the lithologically similar schists of the Antietam Formation (Ꞓ p Ꞓ c in Plate 3). Most of the early workers (for example Stose and Stose, 1944, Figure 26) placed the contact just southeast of the southernmost outcrops of the Conestoga Limestone (O Ꞓ c in Plate 3), but recently Gohn and Thompson (1974) and Wise (1970) have suggested that the line of magnetic highs paralleling this contact a few kilometers to the southeast (Plate 1) is a more likely location.

The aeromagnetic map (Plate 2) provides considerable help in evaluating this problem. Some of the rocks near the top of the Antietam are strongly magnetic and are marked by linear magnetic highs; good examples are 20 km northeast of Hanover, Pa.; 5 km north of Red Lion, Pa., and 5 km north of Quarryville, Pa. (Plate 2). Just northeast of Red Lion, these anomalies continue across the traditional Martic Line into the area originally mapped as Wissahickon. Hietanen (Cloos and Hietanen, 1941, p. 106) reported that the schists associated with these anomalies contain microcline, which is characteristic of the Antietam, but is unknown in Wissahickon schists

elsewhere. Gohn and Thompson (1974) have recently traced the microcline-bearing schists eastward to a point 4 km south of Quarryville, where Cloos (Cloos and Hietanen, 1941) mapped undisputed Antietam to within a few meters of the supposed Wissahickon. For these reasons, we conclude that a thin belt of the schist traditionally mapped as Wissahickon should be assigned to the Antietam. Just southeast of this belt, there are several areas underlain by limestone, including the belt mapped as Conestoga, 6 km east of Hanover, Pennsylvania, and two small patches mapped as Conestoga near Martic Forge, Pennsylvania. Southeast of the limestone, the phyllites are lithologically and magnetically like the porphyroblastic albite-chlorite and paragonite-chlorite phyllites of the Wissahickon in the western Maryland Piedmont (Fisher, 1978). Accordingly, it seems to us that these phyllites belong to the Wissahickon, and we have drawn the contact between the Wissahickon and the lower Paleozoic rocks just southeast of the belt underlain by rock lithologically like the Antietam and Conestoga (Plate 3).

Stratigraphic Relations in the Glenarm Supergroup

The rock types of the Glenarm Supergroup fall naturally into three divisions based on magnetic properties, and lithologic affinities:

A. A lower, essentially nonmagnetic group of rocks, including the Cockeysville Marble and most of the Setters Formation; except the local lenses of garnet-mica schists near the top of the Setters, all of these rocks have magnetic susceptibilities less than 0.10×10^{-4} emu/cm^3.

B. A large group of weakly magnetic, highly aluminous rocks, including the garnet-mica schists of the eastern Wissahickon and the albite-chlorite schists of the western Wissahickon (and its equivalents, the Ijamsville and Urbana phyllites and the "Marburg Schist"); nearly all of these rocks have magnetic susceptibilities in the range 0.5×10^{-4} to 1.0×10^{-4} emu/cm^3. As already discussed, the garnet-mica schists at the base of the Wissahickon in the southeastern Pennsylvania Piedmont are associated with strong magnetic highs and may have high magnetic susceptibility; they are, nevertheless, included in sequence B because of their strong petrographic and stratigraphic affinity with the garnet-mica schists near Baltimore.

C. A large group of strongly magnetic rocks, consisting mainly of quartzose clastic rocks, including the diamictite, metagraywacke, and quartzose schist of the Wissahickon. We have also included in sequence C the strongly magnetic metabasalts of the Sams Creek Formation, which are closely associated with the quartzose mica schists of the Wissahickon in the western Piedmont.

The distribution of these three groups of rocks suggests that each approximates a distinctive time-stratigraphic unit (Figure 15*a*). The critical relations can be seen most clearly in the doubly-plunging anticlinorium culminating in the gneiss anticlines at Baltimore (Plate 3). Here, the rocks of sequence A all crop out at the base of the section and clearly approximate a time-stratigraphic unit at the base of the Glenarm. The weakly magnetic schists of sequence B wrap almost completely around the rocks of sequence A, except in the southeastern part of the anticlinorium, where the section is cut out by the Baltimore Complex. The strongly magnetic, coarse, clastic rocks of sequence C wrap completely around the anticlinorium above the weakly magnetic schists of sequence B, very much as if the contact between sequences B and C approximates a time-stratigraphic boundary.

The same picture seems to emerge from the regional distribution of sequences A, B, and C. The succession in the anticlinorium in the southeastern Pennsylvania Piedmont is virtually identical to that at Baltimore, and the map pattern outlined by sequence C in the saddle between the anticlinorium in Pennsylvania and that at Baltimore is just what we would expect if the rocks of sequence C approximated a single time-stratigraphic unit (Figure 15*a*).

The irregular pattern of anomalies in the northwestern part of the Piedmont (Plate 2) suggests that the structural relations there are more complex, but the same stratigraphic sequences seem to be present. Recent detailed mapping (Fisher, 1978) has shown that the metabasalts of the Sams Creek Formation, which caused the prominent magnetic high trending northeast from Westminster, Md., occupy the keel of a local syncline, and are stratigraphically above the weakly magnetic phyllites of the Wissahickon, which flanks the syncline on the northwest and southeast. Traced northeastward, the metabasalt and the magnetic anomaly it produces gradually approach and finally merge with the magnetic highs associated with the quartzose mica schists of sequence C just north of the Pennsylvania line. Farther southeast, the metabasalt disappears, but scattered

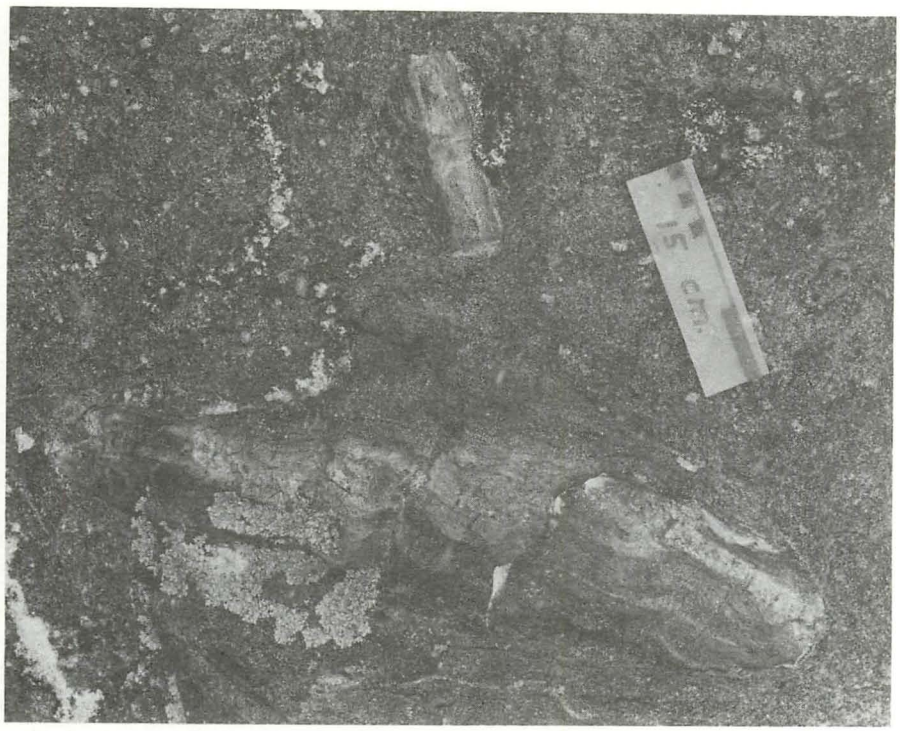


Figure 13. Diamictite of the Wissahickon, showing clasts of metagraywacke in matrix of weakly foliated mica-quartz schist. The light colored parts of the clasts contain abundant epidote and are probably metamorphosed calcareous concretions. Outcrops along the Patuxent River, at State route No. 97, Montgomery County, Maryland.

patches of Wakefield Marble, which is just above the metabasalt in the syncline at Westminster, crop out along the contact between the magnetic rocks of sequence C and the non-magnetic phyllites of sequence B. This map pattern strongly suggests the presence of a northeast-plunging anticline that has the weakly magnetic albite-chlorite phyllites of sequence B in the core and the more magnetic rocks of sequence C—including the Sams Creek Formation, minor Wakefield Marble, and the quartzose Wissahickon schists—on the two limbs and the crest.

The pattern of magnetic anomalies between Westminster and Baltimore suggests that this area is underlain by a set of complexly refolded anticlines bringing up weakly magnetic schists and phyllites of the Wissahickon, alternating with synclines occupied by the strongly magnetic rocks of sequence C. The irregular pattern of these anomalies closely resembles the pattern produced by two interfering sets of folds; Figure 15*b* shows one possible interpretation. Fisher (1978) has mapped fold interference patterns of precisely this kind just west of Westminster, and Higgins (1973) and Freedman, Wise, and Bentley (1964) have described small scale fold interference patterns from other parts of the Piedmont.

Final evaluation of this stratigraphic interpretation must of course await detailed mapping of lithologies and minor structures in the area between Westminster and Baltimore. But our reconnaissance mapping to date has tended to support many features of the distribution outlined in Figure 15*a*. Quartz-rich schists of the Wissahickon do underlie most of the magnetic highs interpreted as synclines in this area, and albite-chlorite phyllites do underlie most of the magnetic lows interpreted as anticlines. In addition, one small outcrop of Wakefield Marble (14 km southeast of Red Lion, Pa.) was described by Stose and Jonas (1939, p. 84); this is the only outcrop of Wakefield Marble in this part of the Piedmont, and it is just along the boundary between strongly magnetic rocks and weakly magnetic rocks, in precisely the same stratigraphic position as the lenses of Wakefield Marble in the syncline near Westminster.

Figure 14. Diamictite of the Wissahickon, showing clast of biotite-quartz gneiss in matrix of weakly foliated garnet-mica-quartz schist; note that the quartz veins and schistosity in the clast stop abruptly at the boundary between clast and matrix, indicating that the clast was metamorphosed prior to deposition in the diamictite. Outcrop along the Patapsco River, 0.5 km east of Sykesville, Carroll County, Maryland.

Sedimentary Facies Relations in the Glenarm

The magnetic anomalies associated with the magnetic garnet schist member of the Setters Formation appear to be confined to the Phoenix and Texas anticlines northwest of Baltimore, suggesting that this member is restricted to the northwestern part of the Setters outcrop belt. Crowley (1976) has confirmed Fisher's (1971) observation that conglomerate lenses in the Setters are restricted to the same area. These relations probably reflect a change in sedimentary facies within the Setters Formation from a conglomerate-quartzite-aluminous shale sequence in the northwestern part of the present outcrop belt to a quartzite-potassic shale sequence in the southeastern part, as was suggested by Fisher (1971).

No changes in the sedimentary facies are discernable on the basis of the available magnetic or geologic data within the Cockeysville Marble or within the weakly magnetic pelites in the lower part of the Wissahickon Group (sequence B of the previous section).

The strongly magnetic quartzose clastic rocks of the upper part of the Wissahickon (sequence C of the previous section) include several rock types, whose distribution strongly suggests a lateral facies relation. Figure 15*c* shows the distribution of rock types in sequence C, already inferred to approximate a single time-stratigraphic unit. Diamictite is restricted to the southeastern side of sequence C; metagraywacke to a strip running northeast-southwest through the central part of sequence C; the quartzose mica schists to a second belt, farther northwest; and the Sams Creek metabasalts to the extreme northwestern part of sequence C. The boundaries between adjacent units of the three metasedimentary rock types, inferred from published geologic maps and our own reconnaissance, are completely gradational

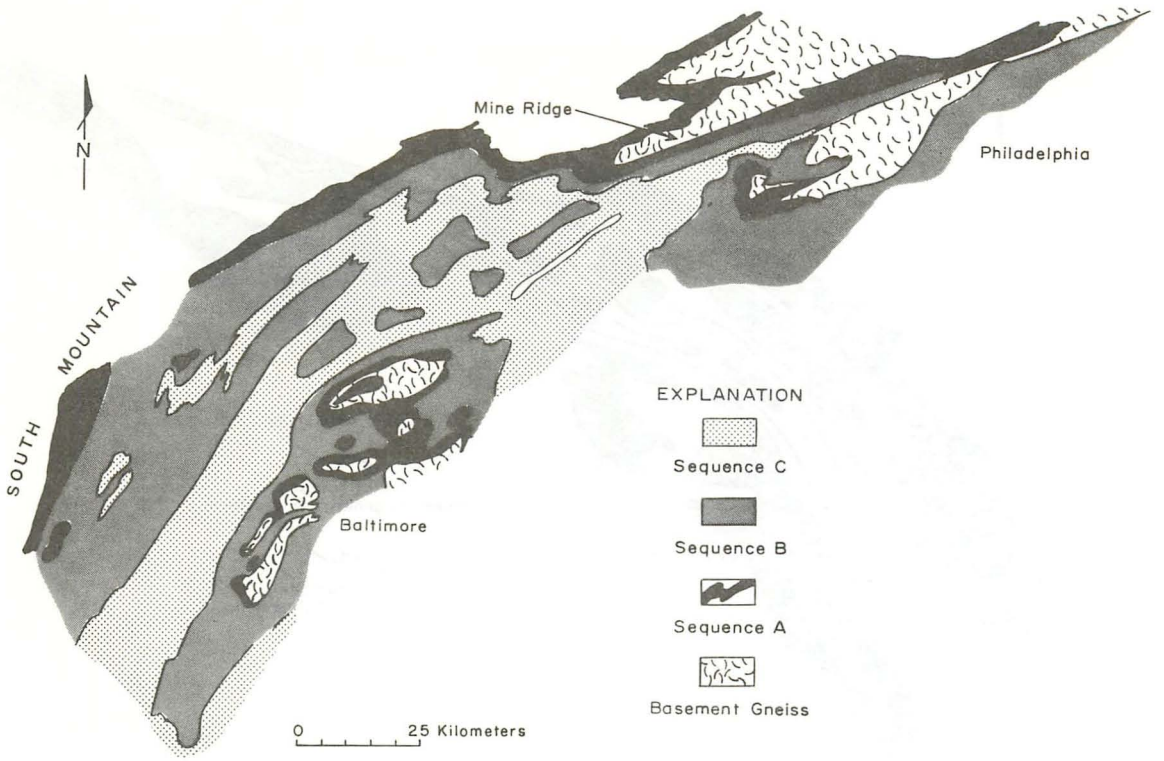


Figure 15*a*. Distribution of major magnetic/lithologic sequences in the northern Piedmont. Sequence C consists of quartzose metasedimentary rocks, strongly magnetic for the most part, including the Wissahickon diamictite, metagraywacke, and quartz schist, and the Sams Creek Formation. Sequence B consists of pelitic rocks, weakly magnetic for the most part, including the Wissahickon pelitic schists, the Urbana and Ijamsville phyllites, and the "Marburg" and "Octoraro Schists" (of local usage). Sequence A consists of nearly nonmagnetic carbonate rocks and quartzites, including the Setters Formation and Cockeysville Marble near Baltimore and Philadelphia; the Chickies, Weverton, Harpers, Antietam, Frederick, and Grove Formations; and the Beekmantown and Conococheague Groups near South Mountain and Mine Ridge.

and suggest to us that these three rock types represent lateral facies equivalents of one another.

This gradation is best seen in the anticlinorium centering on the Baltimore Gneiss anticlines, where a continuous transition from one rock type to another can be seen by following the contact between the magnetic rocks of sequence C and the weakly magnetic rocks of sequence B from one side of the anticlinorium to the other. On the southeast, the rocks near the base of sequence C consist exclusively of coarse, bouldery diamictite, containing cobbles and boulders of various rock types, slabs of contorted sediments, and pebbles of vein quartz, scattered about in a completely unbedded, structureless matrix having the composition of a coarse, sandy graywacke (Figs. 13 and 14). This lithology can be followed around the southern nose of the anticlinorium, then back to the northeast along the western limb of the anticlinorium, as far as Sykesville, Maryland (Plate 3). Virtually identical rocks crop out on the northeastern side of the anticlinorium in the belt of diamictite near Fallston, in Harford County, Maryland.

West of Sykesville and Fallston, the largest clasts in the diamictite decrease in size from boulders to cobbles and finally to pebbles, and scattered interbeds of coarse, graded graywackes as much as 1 m thick appear, interbedded with diamictite. Farther northwest, the proportion of metagraywacke increases, and the proportion of diamictite is correspondingly reduced. Simultaneously, the bedding in the metagraywackes thins to about 20 cm, maximum grain size decreases to 0.5 cm, mica schist becomes thinly interbedded with metagraywacke, and the meta-

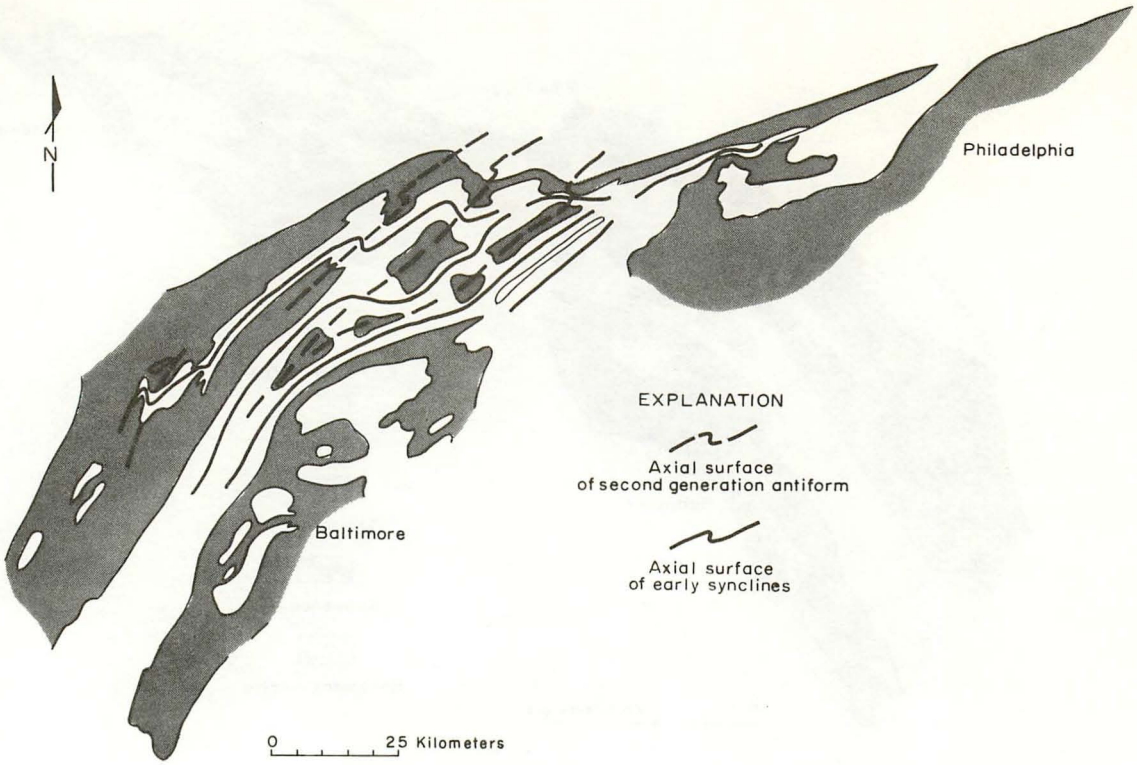


Figure 15*b*. Schematic tectonic map illustrating interpretation that the distribution of sequences B and C in the Wissahickon reflect interference of two fold systems; area shown is the same as Figure 15*a*.

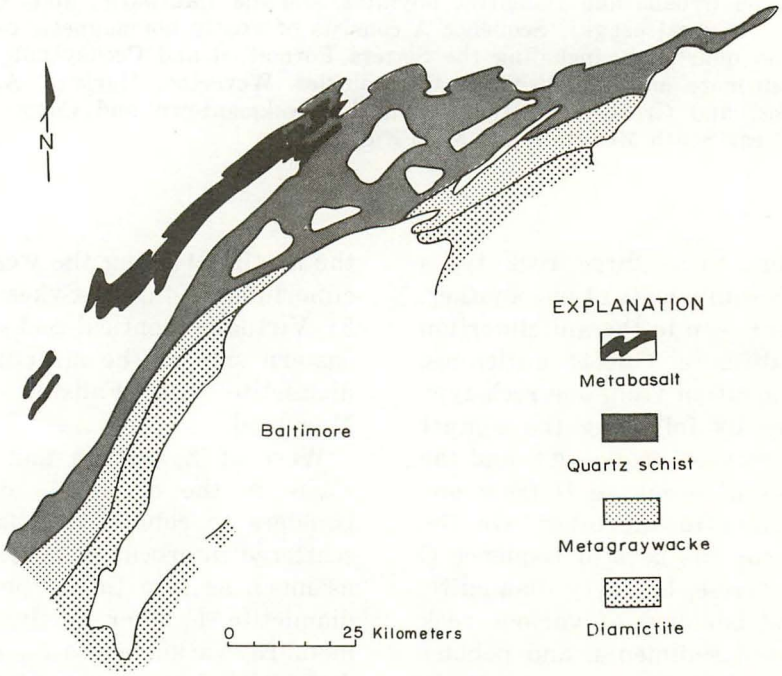


Figure 15*c*. Map showing generalized distribution of rock types in sequence C of Wissahickon; contacts between lithologies are highly gradational, and are only approximately located in the northeastern part of the map area. The overall distribution of sediment types suggests a southeasterly source. Area shown is the same as Figure 15*a*.

graywackes have the internal structure of turbidite deposits (Fig. 12 and Fisher, 1970). In Figure 15*c*, we have arbitrarily drawn the boundary between metagraywacke and diamictite at the point where metagraywacke appears to predominate, but the contact between the two is highly gradational. The transition is best displayed at the northern tip of the anticlinorium, 10-15 km north of Fallston, Md., in the area near Rocks State Park (*see* Southwick and Owens, 1968). It is less well displayed near Reisterstown on the northwestern flank of the anticlinorium due to the scarcity of outcrop, and probably because of extreme shortening of the section by folding, suggested by the crowding of magnetic anomalies in this area (Plate 2). Nevertheless, Crowley (1976) has mapped a lens of metagraywacke (his Piney Run Formation) as being just north of the diamictite (his Sykesville Formation), and interfingering with the diamictite.

Still farther north and west, near White Hall, Maryland, the metagraywackes become thinner bedded, finer grained, and more delicately laminated, and closely resemble distal turbidite deposits (Fig. 11). Similar rocks crop out throughout the area occupied by sequence C north and west of White Hall; the lithology changes little except that the proportion of clearly recognizable bedding gradually diminishes to the northwest.

Exceptions to the relations described above do exist: for example, small lenses of diamictite are present in the metagraywackes as far west as Great Falls, Maryland, and Crowley (1976) has described metagraywacke at the base of sequence C in the vicinity of Baltimore City. However, in general, the distribution of rock types in sequence C appears to record a systematic gradational change from diamictite on the southeast through metagraywacke to quartzose mica schists on the northwest. And if these rocks are all part of a single time-stratigraphic unit, as inferred above, then this transition must record a lateral change in original sedimentary facies.

Interpretation of Sedimentary Environments

The diverse lithologies of the rocks in sequences A, B, and C suggest deposition in different sedimentary environments. Hopson (1964), Southwick (1969), Higgins (1972), and Crowley (1976) have all interpreted the Setters Formation and the Cockeysville Marble as shelf sediments, representing deposition of clean quartz sand, potassic mud, and carbonate sediments during fairly stable tectonic conditions, presumably in shallow water, possibly under conditions of restricted circulation. To this interpretation we can add only the thought that the local conglomerate lenses in the northwest part of the Setters outcrop belt may imply a northwesterly source.

The uniform, highly aluminous composition of the weakly magnetic schist and phyllite of sequence B suggests that these rocks reflect pelagic sedimentation in an extensive basin, presumably marine, developed on top of the preexisting carbonate bank.

The distinctive, well-preserved sedimentary structures and composition of the Wissahickon diamictite and metagraywacke have led all recent investigators to interpret the diamictite facies as a complex of overlapping submarine slide deposits and the metagraywacke facies as an extensive submarine turbidite fan (Hopson, 1964; Fisher, 1970; Southwick, 1969; and Higgins, 1972). However, the uncertainties regarding the structure and stratigraphy of the Wissahickon have led to numerous difficulties in interpreting the relations between the diamictite, the metagraywacke, and the other rocks in the Wissahickon.

Recognition of the probable time equivalence of the rocks in sequence C of the previous section suggests that the diamictite, the metagraywacke and the quartzose mica schist facies of the western part of sequence C constitute a coherent group of consanguineous flysch sediments. The diamictite facies on the southeast probably represents coarse, bouldery debris flow deposits in a complex of coalescing submarine fans. These fans appear to interfinger with and grade northwestward into an apron of turbidite deposits, now represented by the metagraywacke, deposited by turbidity currents spreading outward towards the northwest from the foot of the fan complex. The delicately laminated quartzose mica schists closely resemble distal turbidite deposits, and probably represent the northwesternmost part of the turbidite apron.

If this picture is correct, the metagraywacke at the base of the diamictite near Baltimore could reflect clastic sediments shed into the basin before the main submarine fan complex was fully established. The small pockets of diamictite interbedded with metagraywacke near Great Falls, Maryland, could represent unusually far-travelled debris flow slides from the main part of the fan complex, or possibly, local slumps generated on the turbidite apron itself, well beyond the fan complex.

The Sams Creek metabasalts appear at the very base of the quartzose mica schists that character-

ize the westernmost part of the flysch sequence, and appear to reflect a chain of volcanic centers that were active in the western Piedmont just prior to the beginning of flysch sedimentation.

The arrangement of sedimentary facies in the flysch sequence clearly indicates that the source for these sediments lay to the southeast. The nature of this source area can best be inferred from the composition of the clasts in the diamictite. As discussed by Hopson (1964) and Fisher (1970), many of the clasts are fragments of mica schist and contorted slabs of metagraywacke probably derived by slumping and redeposition of sediments within the basin (Fgi. 13). The clasts which appear to be derived from outside the basin include pebbles of vein quartz, cobbles and boulders of amphibolite, and scattered cobbles of biotite-quartz gneiss containing quartz veins and foliation that clearly predate deposition of the clasts (Fig. 14). Hopson (1964) and Fisher (1970) interpreted the biotite-quartz gneiss clasts as Baltimore Gneiss, and Hopson (1964, p. 131) therefore inferred that the Wissahickon sedimentary rocks were derived from crystalline rocks like those in the present basement.

TABLE 4

Modal analyses of felsic gneiss clasts from Wissahickon diamictite and comparison with analyses of James Run Formation

	(1)	(2)	(3)	
			Average	Std. Dev.
Plagioclase	46.2	49.2	46.0	4.6
Quartz	36.2	35.9	41.3	1.3
Microcline	—	0.6	0.6	1.2
Hornblende	—	—	4.2	4.5
Biotite	13.0	11.2	2.9	1.3
Muscovite	tr	0.1	2.2	4.2
Epidote	1.2	1.0	1.5	0.8
Opaques	2.4	2.0	0.7	0.3
Sphene	0.2	tr	tr	—
Zircon	tr	tr	0.1	0.1
Apatite	0.2	tr	tr	—
Allanite	tr	tr	tr	—
Garnet	0.1	tr	0.5	—
Total	100.0	100.0	100.0	
Pts. Counted	909	716	—	

1) Felsic gneiss clast in diamictite (1802-A); Sykesville, Md.

2) Felsic gneiss clast in diamictite (1801-F); Sykesville, Md.

3) Average and standard deviation of five samples of felsic gneiss from the James Run Formation, Harford County, Md. (Southwick, 1969, Table 5)

However, a closer look at these clasts shows that most are quite unlike the basement gneisses that are presently exposed. We have studied approximately 30 thin sections of these clasts, and most are garnet-biotite-quartz-andesine gneiss; some contain hornblende instead of biotite, but none contain more than 1-2 percent of potassic feldspar (Table 4). In contrast, the Baltimore Gneiss characteristically contains 25-35 percent potassic feldspar, 10-30 percent sodic plagioclase, and virtually no garnet (*see* modal analyses in Hopson, 1964). In addition, most of the basement rocks are coarsely segregated, thoroughly migmatized rocks, which clearly have been subjected to prolonged, high-grade metamorphism (Olsen, 1972). Most of the metamorphic clasts in the diamictite, however, have textures more reminiscent of low-grade metamorphic rocks; the only segregations are thin biotitic laminae parallel to the cleavage, the rock is generally fine grained, and the veins cutting the cleavage are nearly pure quartz, quite unlike the feldspar-rich migmatitic veinlets of the Baltimore Gneiss. For these reasons, the Baltimore Gneiss does not appear to be the source of these clasts.

The only rocks now exposed in the Maryland Piedmont which resemble these clasts are the felsic and mafic volcanic rocks included in the James Run Formation (Southwick, 1969; Higgins, 1972; Crowley, 1976). Compositionally, the felsic volcanic rocks are very similar to the clasts in diamictite (Table 4; *see* also modal analyses in Southwick, 1969, p. 48). None of the clasts we have found so far show the interbedding of mafic and felsic rock types that is so common in the James Run; but not of all the James Run contains layering thin enough to show in the clasts (M. W. Higgins, unpublished data, 1974). A part of the formation (the upper part, according to Southwick's interpretation, 1969, p. 46) contains layers of felsic gneiss as thick as 6 m. Cobbles from these beds would closely resemble the biotite-quartz gneiss clasts in the diamictite. Furthermore, the James Run and related volcanic rocks formerly assigned to the Baltimore Complex are exposed in an extensive belt just southeast of the diamictite (Plate 3), precisely the direction from which the sediments in the diamictite came. Finally, our interpretation that the gneissic clasts are derived from the James Run meshes perfectly with Crowley's (1976) observation that the Wissahickon metasedimentary rocks associated with the diamictite contain blocks and slabs of serpentinite from the Baltimore Complex, which is closely associated with the James Run Formation, just southeast of the diamictite (Plate 3).

Crowley (1976) interpreted the Baltimore Complex and the James Run metavolcanic rocks as one or more fault slices, thrust into the Wissahickon basin during sedimentation. All of our observations support this conclusion. Crowley's mapping and our own reconnaissance clearly indicate that the contact between the Baltimore Complex and the underlying Wissahickon metasedimentary rocks is indeed a thrust fault, because it cuts major lithologic units within the Baltimore Mafic Complex as well as within the Wissahickon. This relation is particularly clear in the part of the complex between Baltimore and Conowingo, Maryland (Plate 3). Traced southward from Conowingo, the thrust cuts progressively up section from ultramafic rocks through gabbro to laminated metavolcanic rocks in the Baltimore Mafic Complex (the hanging wall) at the same time as it cuts down section from diamictite through the Wissahickon pelite to Cockeysville Marble in the footwall. Similar relations are shown by Crowley's (1976) mapping in western Baltimore City. And, because clasts in the diamictite are apparently derived from the Baltimore Complex and James Run metavolcanic rocks, thrusting must have taken place during sedimentation. Therefore, the clastic sequence in the upper part of the Wissahickon presumably represents a flysch fan deposited in front of the advancing thrust block. The flysch fan is composed partly of sediments derived from the thrust block itself and partly of previously deposited Wissahickon sedimentary rocks, ripped up and redeposited in front of the advancing thrust block, much as snow is redeposited in front of a moving snow plow. This interpretation provides a ready explanation for the distribution of facies in the Wissahickon (Fig. 15), and for the immense thickness of the diamictite. Deposition of a slide deposit as much as 3,000 m thick, and 200 km wide along strike without a single discernable interruption by bedded sediments, must reflect a supremely dramatic event; and surely the emplacement of a thrust block the size of the Baltimore Complex would qualify!

To summarize, the lowermost rocks of the Glenarm (Setters Formation and Cockeysville Marble) appear to represent a shelf sequence, deposited on a stable continental platform; the lower pelitic part of the Wissahickon appears to represent a sequence of pelagic rocks, deposited in a deepening marine basin; and the diamictite, metagraywacke, and quartzose pelitic schist in the upper part of the Wissahickon appears to represent a major sequence of flysch sediments derived from a southeasterly source, probably triggered by thrusting of the Baltimore Complex and the James Run Formation into the basin during sedimentation. The rocks deposited in these three different environments are generally characterized by distinctive magnetic properties, which should help in tracing these rock units farther south into the Virginia Piedmont.

Age and Correlation of the Glenarm

Our reconstruction of regional stratigraphic and structural relations in the northern Piedmont has brought to light several points that bear critically on the long-standing controversy regarding the age and correlation of sequences in the Glenarm Supergroup (*see* reviews in Higgins, 1972; and Seiders and others, 1975).

First is the close relation between the Setters Formation and the basal Paleozoic clastic sequence in the South Mountain and Mine Ridge anticlinoria (Fig. 15*a* and Plate 3). In the anticlinoria bounding the Piedmont on the northwest, the basal Paleozoic section (Єp Є c in Plate 3) includes local lenses of quartz pebble conglomerate near the bottom (Hellam member of Chickies Quartzite), succeeded upward by feldspathic quartzite (Weverton Quartzite in Maryland, Chickies Quartzite in Pennsylvania), potassic shale (Harpers Formation) and, at the top, a second quartzite unit (Antietam Quartzite). Followed northeastward along the anticlinoria, the Antietam becomes shalier, and at Mine Ridge can no longer be distinguished from the Harpers. An Early Cambrian age for the upper part of this sequence has long been accepted on the basis of Early Cambrian fauna found in the upper part of the Antietam in Virginia, Maryland, and Pennsylvania (e.g. Walcott, 1892; Stose and Stose, 1946, p. 42; Stose and Stose, 1944, p. 15).

The Chickies and the Harpers-Antietam sequence can be followed northeastward from Mine Ridge in a nearly continuous outcrop belt along the southeast side of the Mine Ridge anticline, to a point near Norristown, Pennsylvania, where the Chickies swings southwestward in a southwest trending syncline (Plate 3). The southeast limb of the syncline can be followed southwestward only a short distance before it is cut off by a major fault along the northwest side of the West Chester anticline; however, the Setters Formation crops out in precisely the same stratigraphic position in the Woodville anticline, just 40 km to the southwest (Plate 3). Furthermore, the Setters has many points in common with the basal Paleozoic clastic

sequence. In the northernmost gneiss anticlines in Maryland (Phoenix anticline), the Setters has precisely the same stratigraphic succession as in the basal Paleozoic section exposed directly across the synclinorium to the northwest: local lenses of quartz pebble conglomerate at the base, feldspathic quartzite above, and highly aluminous pelite at the top (Fisher, 1971). The basal conglomerate seems to disappear to the southeast, and the pelite becomes less aluminous, and more potassic. But the general lithologic character of the Setters remains very similar to the basal clastic section in Mine Ridge; both are characterized by unusually high K_2O content and by abundance of tourmaline. They even have the same magnetic expression: both are marked by nearly flat magnetic anomalies of low amplitude and have local magnetic highs that reflect lenses of magnetite-rich pelite near the top. The similarity of the pattern produced by the Chickies and the Antietam near Quarryville, Pennsylvania, to that produced by the Setters Formation rimming the Phoenix anticline is especially striking.

In view of the near continuity of this group of clastic rocks and their strong lithologic similarity, we conclude that the Setters Formation is simply the southeastward extension of the basal Paleozoic clastic sequence.

This conclusion is further supported by the relations in the carbonate rocks. The fossiliferous carbonate rocks that overlie the Chickies and the Harpers near Lancaster, Pennsylvania, can also be traced northeastward, along the south rim of the Mine Ridge anticline and around the hinge of the southwest plunging syncline at Norristown; they too are cut off by the fault bounding the northwest side of the West Chester anticline but seem to reappear as the Cockeysville Marble above the Setters Formation in the Woodville anticline 40 km southwest of the fault (Plate 3). Once again, a correlation seems almost inevitable. Stratigraphic relations within the Cockeysville are complex (Crowley 1976) and have been little studied; however a unit of phlogopitic calcite marble (Crowley's phlogopite metalimestone member) is common at the top of the Cockeysville near Baltimore, and L. M. Hall (oral communication, 1972) has commented that the unit is strongly reminiscent of the carbonate rocks at the base of the Manhattan Schist in southeastern New York, which contain Early Paleozoic pelmatazoan columnals, and have been correlated with the Cambrian and Ordovician carbonate section in Vermont (Hall, 1968).

The distribution of the pelite and flysch sequences in the Wissahickon (Fig. 15*a*) strongly suggests that the Wissahickon is at least broadly conformable with the Cockeysville and the Setters. Some authors have suggested that part of the Wissahickon may be allochthonous, as a thrust block or series of thrust blocks, perhaps analogous to those in the Taconics (Kay, 1941; Rodgers, 1970) or in the Smith River allochthon of Conley and Henika (1973) as suggested by Rankin (1975). These possibilities are very difficult to prove or disprove in the absence of fossils, but we find no compelling evidence for major thrusting of this sort within the Wissahickon.

The configuration of rock types in the Wissahickon that we have inferred (Figs. 15*a*, *b* and *c* and Plate 3) places some constraints on any such proposals. The most striking characteristic of the flysch sequence in the upper part of the Wissahickon is its internal coherence, reflected both by the sequence of facies (Fig. 15*c*) and by the smooth regional variation in grain size and bedding thickness; local faults are present (Southwick, 1969; Drake, written communication, 1977) but there is no indication of major internal shuffling of the parts of the flysch sequence. Rankin (1975) proposed a regional thrust fault at the contact between the flysch sequence and the underlying pelitic rocks; local faulting is present at the base of the Sams Creek Formation near Westminster, Maryland (Fisher, 1978), but once again several lines of reasoning militate against a regional thrust fault at this contact. First, the thrust at the base of the Baltimore Complex clearly cuts the contact between the flysch sequence and the underlying pelite (Plate 3, and Crowley, 1976), so any thrust at the flysch-pelite contact must predate emplacement of the Baltimore Complex. But this sequence is incompatible with our conclusion that emplacement of the Baltimore Complex triggered deposition of the flysch sediments. Second, isolated slide blocks of serpentinite and lenses of diamictite composed largely of serpentinite are found within the pelitic rocks well below the flysch sequence. These rocks appear to be precursors of the thrusting that emplaced the Baltimore Complex, and suggest a close genetic relationship between the supposedly autochthonous pelitic rocks and the flysch sequence.

Thrusts could occur within or at the base of the pelitic sequence. This possibility is at the heart of the classic interpretation of the Martic Line, and both Kay (1941) and Rodgers (1970) have proposed faults at this point in the section. Minor faulting may well be present at the carbonate-schist contact and may have contributed to the

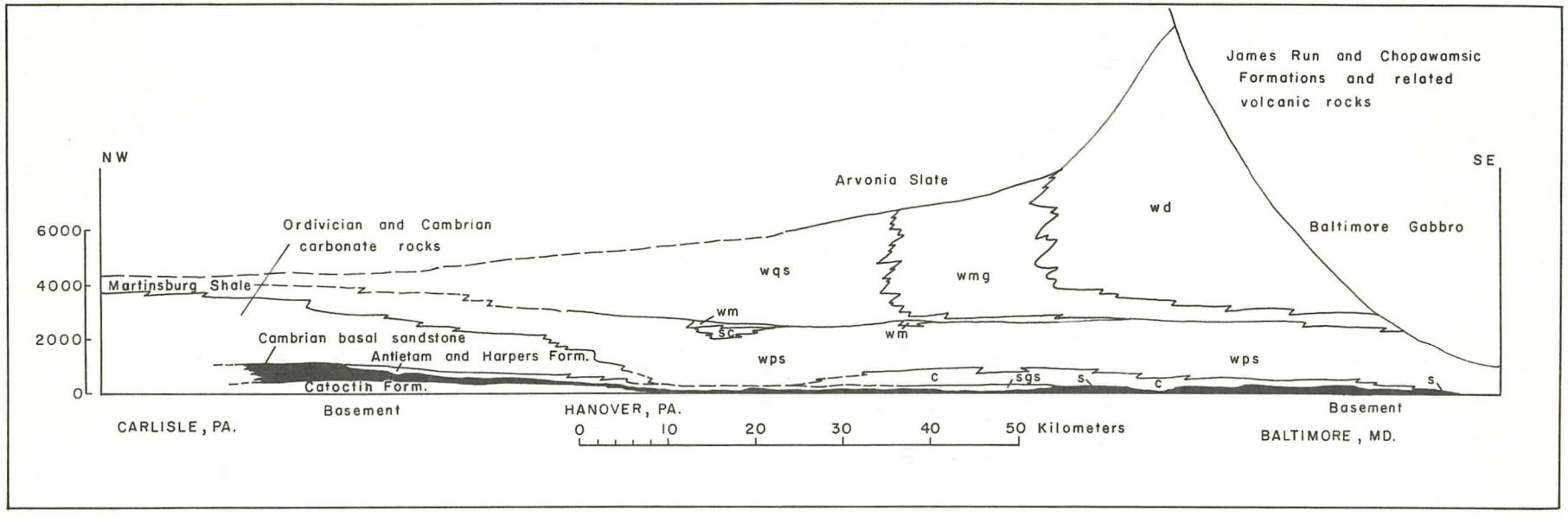


Figure 16. Restored stratigraphic cross section of the northern Piedmont from Baltimore, Md. through Hanover to Carlisle, Pa., showing the inferred relation between Piedmont rock units and Lower Paleozoic rocks. Shortening due to the second fold generation in the Piedmont has been removed, assuming that folds of this generation were concentric. The locations of time lines in this cross section are uncertain; see text for discussion. Symbols as in Plate 3.

complexities of the relations south of Quarryville, Pennsylvania. However, there is no evidence of any major stratigraphic offset at this contact. Because stratigraphic units in the Wissahickon appear to conform with those beneath it (Fig. 15*a*, Plate 3), any thrust at the base of the Wissahickon must nearly parallel bedding throughout the sequence, without cutting across stratigraphic units in either footwall or hanging wall. Although possible, such behavior is most uncommon for a thrust of the required size.

To sum up, it is intriguing to speculate about possible regional thrusting within the Glenarm, but serious objections exist to most of the locations for such a fault that have been suggested so far. At this time, we see no convincing evidence that regional faulting is required and conclude that the Wissahickon could be in normal stratigraphic position on the Cockeysville Marble.

Although no one point of evidence that we have cited constitutes proof, we feel that the weight of evidence strongly favors a correlation between the Glenarm and the upper Precambrian and lower Paleozoic section exposed in the South Mountain and Mine Ridge anticlines (as proposed by Miller, 1935), rather than a correlation of the Glenarm with the upper Precambrian sequence of the southern Appalachians (as proposed by Hopson, 1964) or with part of an older Precambrian sequence of metamorphic rocks.

Our interpretations of these correlations are summarized in Figure 16, drawn along a line stretching north-northwest across the Piedmont from Baltimore through Hanover to Carlisle, Pennsylvania. Note that we have not connected the Cockeysville directly to the Cambrian and Ordovician carbonate bank to the northwest. We suspect that a connection between the two existed in the area between Norristown, Pennsylvania, and the Woodville anticline, but Rodgers (1968) and Reinhardt (1974) have made a strong case for the disappearance of the carbonate bank a few kilometers or tens of kilometers southeast of a line connecting Frederick, Maryland, and Lancaster, Pennsylvania; probably a deep embayment existed in the carbonate bank between Baltimore and Frederick, extending much of the way to Norristown, somewhat like the embayment in the present Bahama bank at the Tongue of the Ocean.

Even assuming that the general correlation scheme shown in Figure 16 is correct, however, there is still room for debate regarding exact ages of many of the rocks in the Glenarm. The only unequivocal fossil dates so far available in the northern Piedmont are from the Arvonia Slate, which uncomformably overlies metavolcanic and metasedimentary rocks equivalent to the Wissahickon in a small syncline in the central Virginia Piedmont. The Arvonia contains Late Ordovician fossils (Stose and Stose, 1948) and the Glenarm Supergroup must therefore be at least that old. If the correlation in Figure 16 is correct, it must also be younger than the Catoctin Formation in the Blue Ridge, dated as 820 m.y by Rankin and others (1969).

No fossils have been found in the rocks assigned to the Glenarm. However, Early Cambrian fossils have been recovered from the uppermost beds of the Antietam Quartzite in Virginia, Maryland, and Pennsylvania (Walcott, 1892, p. 52-57; Stose and Stose, 1944, p. 15; 1946, p. 42), and if the upper part of the Setters correlates with the Antietam, the Setters probably ranges in age from latest Precambrian at the base to at least Early Cambrian at the top. It could be still younger, however; in the Frederick Valley, just 70 km west of Baltimore, the upper Precambrian and lower Paleozoic clastic sequence persisted until Middle Cambrian time (Reinhardt, 1974). The base of the Cockeysville is presumably younger than the Setters, possibly Early or Middle Cambrian. As discussed above, the phlogopite-calcite marble near the top of the Cockeysville resembles the basal carbonate beds of the Manhattan Schist, which has been correlated with Middle Ordovician carbonate rocks in Vermont, and with limestone containing pelmatazoan columnals at Verplanck Point, New York (Hall, 1968; Ratcliffe and Knowles, 1969). However, it is a long way from Baltimore to Vermont, and the age of the phlogopite-calcite marble in the Cockeysville could differ substantially from that of the rocks in Vermont.

Numerous radiometric dates have been measured in the Piedmont, but the results are not entirely consistent (*see* Seiders and others, 1975; Higgins and others, 1977); apparently the method does not yet have the resolution to pinpoint stratigraphic ages in the Piedmont. All that can be said with confidence is that *if* our interpretation that the cobbles of biotite-quartz-andesine schist in the Wissahickon diamictite are derived from the James Run Formation is correct, this part of the diamictite must be younger than the James Run, which has yielded lower Paleozoic zircons (Tilton, Doe, and Hopson, 1970; Higgins and others, 1977). On the other hand, it may not be much younger.

In summary, the lithologic correlations outlined in Figure 16 seem reasonably reliable to us, but the locations of time lines within this section, and in particular the time relations between the Wissahickon Group and the Martinsburg Shale, are still uncertain. The carbonate bank at Baltimore may

have been flooded well before the bank farther west, and much of the Wissahickon may be older than the Martinsburg. Alternatively, if the upper part of the Cockeysville is equivalent in age to the rocks at the base of the Manhattan Schist and the Middle Ordovician section of Vermont (Hall, 1968), and if the Wissahickon is autochthonous, the entire Wissahickon must have been deposited during the Middle Ordovician, roughly 30 m.y. long. Sedimentation rates of the order of 400 m/m.y. would be required, but rates of this magnitude are common in the Cenozoic, where nearly complete sedimentary sections have been measured (Pettijohn, 1975, p. 595), and appear reasonable for the Wissahickon sediment types. Or, if a part of the Wissahickon is allochthonous, the flysch sequences could have been deposited at any time during the 100 m.y. between eruption of the James Run volcanic rocks (about 550 m.y. ago on the basis of zircon data from Tilton and others, 1970) and deposition of the Arvonia Slate (Middle to Upper Ordovician, about 450 m.y ago).

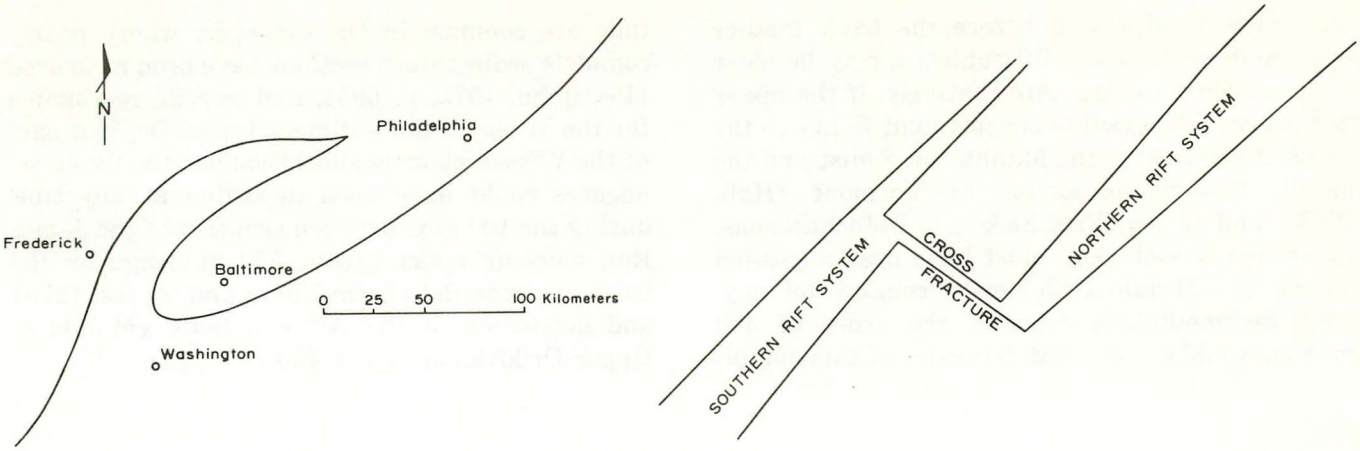
PLATE TECTONICS AND THE NORTHERN PIEDMONT

Numerous authors have attempted to interpret Appalachian evolution in terms of plate models (for example, Bird and Dewey, 1970; Glover and Sinha, 1973; Odom and Fullagar, 1973; and Rankin, 1975). Most recent models have invoked four stages of evolution; (1) an initial period of late Precambrian continental rifting, recorded by the deposition of clastic sediments such as the Ocoee Supergroup and eruption of volcanic rocks including those of the Catoctin Formation; (2) development of an Atlantic-type shoreline on the rifted continental margin, recorded by the upper Precambrian and lower Paleozoic quartzite-carbonate sequence; (3) development of an island arc east of the Paleozoic margin of North America, and the onset of plate convergence, recorded by volcanic sequences along the eastern edge of the Piedmont; and (4) continental collision, leading to telescoping of sequences, folding, uplift, and deposition of the upper Paleozoic clastic sequences in the Valley and Ridge province. Despite this broad agreement, many details have proven controversial, especially the original position and orientation of the island arc systems developed during plate convergence. The geologic history which we have inferred for the northern Piedmont in the previous sections is clearly compatible with the broad outlines of this evolutionary sequence, but several points bearing on the details of the model deserve mention.

The area discussed in this report straddles two of the sharpest bends in the Appalachians. Rankin (1976) has suggested that this configuration is a reflection of the original pattern of late Precambrian rifting, and that each of the sharp bends in the Appalachians represents a fossil triple junction of the sort visualized by Dewey, and Burke (1974). However, the configuration we have inferred for the Cambrian and Ordovician carbonate bank suggests an intriguing alternative. As already noted, a deep embayment appears to have existed in the carbonate bank northwest of Baltimore (Fig. 17*a*), suggesting that the shape of the shoreline in Cambrian and Ordovician time may have been more complex than implied by the gross configuration of the Appalachians at this latitude.

The literature on rift geometry (Baker, Mohr, and Williams, 1972), on the geology of rifted continental margins (Campos, Ponte, and Miura, 1974), and on rifting mechanisms (Oldenburg and Brune, 1975) all suggest that the detailed shape of rifted continental margins may be more complex than implied by simple coalescence of three-armed rift systems, radiating from a string of triple junctions developed over domical uplifts. Rifts from adjacent uplifts are commonly parallel to one another because of control by pre-existing structural features, but they are rarely exactly co-linear. Consequently, as rift systems propagate toward one another from adjacent uplifts, they tend to pass one another by rather than meeting tip to tip, and connections between two rift systems often form by growth of short cross fractures, originating at the tip of one rift system. Many such cross fractures ultimately develop into transform faults (Oldenburg and Brune, 1975). The resulting geometry (Fig. 17*b*) closely approximates the inferred shoreline near Baltimore, especially if allowance is made for later deformation during continental collision.

The distribution and thickness of upper Precambrian and lower Cambrian rocks are in accord with this interpretation. In the southern Appalachians, an immensely thick sequence of upper



(*a*) (*b*)

Figure 17. Sketch map illustrating the interpretation that the generalized shape of the carbonate bank edge (Figure 17*a*) could be inherited from the geometry of late Precambrian rift systems that initiated the opening of the early Paleozoic ocean east of North America (Figure 17*b*). See text for discussion. The area shown and map scale are the same in both parts of the figure.

Precambrian clastic rocks (the Ocoee Supergroup, 12,000 m thick, according to Hadley, 1970) is overlain by a thick Cambrian clastic sequence (the Chilhowee, up to 2,000 m thick according to Schwab, 1972). Volcanic rocks (the Catoctin Formation) commonly are present near the contact between the two clastic sequences. The thicknesses vary somewhat irregularly (e.g. Schwab, 1972, Fig. 3), but traced northeastward along strike, the sequences thin, and eventually pinch out. The upper Precambrian clastic rocks disappear in northern Virginia, the volcanic rocks disappear in southern Pennsylvania, and although the lower Cambrian sequence can be traced all the way to New York City, it is only a meter or two thick there. Similarly, in the northern Appalachians, a comparable upper Precambrian sequence, 1,000 m thick in Vermont, thins southward to virtually nothing at New York City (Rodgers, 1972). This distribution suggests to us the existence of two distinct rift systems: one, emanating from a triple junction near Montreal (Burke and Dewey, 1973), and another emanating from the southern Appalachians, possibly near Mount Rogers (Rankin, 1976). The two rift systems are characterized by distinct sedimentary and volcanic sequences during their early stages, and both sequences are thinner and younger toward the distal ends of the inferred rifts. Therefore the two rift systems may have originated independently, and propagated toward one another during Late Precambrian time. The sedimentary fill in the southern Appalachian rift system is so much thicker than that in the northern that it is tempting to infer that the southern system began to form earlier.

The two systems appear to have coalesced by Early Cambrian time, when the alluvial and deltaic sediments in the upper part of the Chilhowee (Schwab, 1972) were deposited. But stratigraphic relations near Baltimore suggest that the rift system was still expanding just prior to deposition of these rocks. As pointed out above, the distribution of conglomerate in the lower part of the Setters Formation implies a northwesterly source, as in the basal rocks of the Chilhowee to the northwest. If correct, this implies that the Setters was a part of the same alluvial-fan and flood-plain complex as the rocks farther northwest, and that the embayment inferred to have been present during growth of the carbonate bank had not yet isolated Baltimore from the craton to the northwest. The later development of the embayment could reflect the final propagation of the southern rift system, just prior to coalescence of the two systems.

Development of the nearly continuous Cambrian and Ordovician carbonate bank above the Chilhowee presumably reflects the establishment of a continuous shoreline on the eastern margin of North America. Flooding of this bank by Wissahickon pelite reflects subsidence of the eastern edge of the bank. Fisher (1978) has mapped microcline-bearing quartzite interbedded with the pelite of the lower part of the Wissahickon, requiring a nearby source of crystalline rocks, possibly local horsts within the basin, the last vestiges of the rift system.

The James Run volcanic rocks may represent volcanism in an island arc or associated marginal basin environment (Crowley, 1976, p. 13; Higgins,

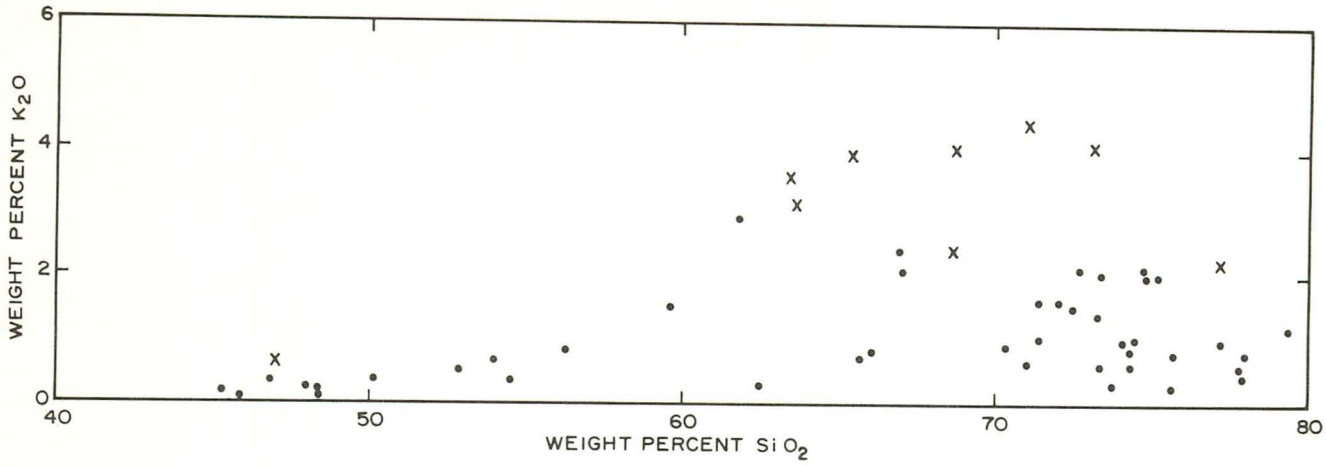


Figure 18. Plot of weight percent K_2O versus weight percent SiO_2 for northern Piedmont igneous rocks. The dots represent analyses of the James Run Formation, the Baltimore Complex, and other igneous rocks from the same thrust block; the crosses represent analyses of igneous rocks which intrude Glenarm metasedimentary rocks. Data from Higgins (1972 and unpublished data), Hopson (1964), and Southwick (1969).

1972, p. 1001). According to D. L. Southwick (written communication, 1974), the plutonic part of the associated Baltimore Complex has a bulk composition suggestive of shallow sub-arc magma reservoirs. The entire complex of plutonic and volcanic rocks may represent the remnants of an island arc developed somewhere southeast of its present location. If the plutonic rocks cutting the Glenarm were produced by melting along the same Benioff zone, the zone must have dipped west, beneath the North American continent, because all the volcanic rocks apparently were derived from some distance to the southeast, whereas at least some of the plutonic rocks intrude clearly autochthonous rocks (including the Baltimore Gneiss, the Setters Formation, and the Cockeysville Marble) and must have been intruded in their present locations, well northwest of the original arc. Available chemical data are fully compatible with this intrepretation; despite some overlap, the rocks which intrude the Glenarm are distinctly more potassic than those in the thrust block containing the Baltimore Complex (Fig. 18), just as would be expected if the intrusions were emplaced well to the continental side of an arc (Dickinson, 1970).

Glover and Sinha (1973) have suggested that the Carolina slate belt records an island arc developed on the eastern side of the Paleozoic Atlantic Ocean, dipping beneath Africa. Dewey (1969) suggested a similar interpretation for the Avalon belt in Newfoundland. Recently, several authors (for example Rankin, 1975) have suggested that the James Run volcanic rocks could be a part of this same belt. However, several lines of reasoning suggest that the James Run is not part of the slate belt volcanic sequence. The James Run contains none of the sedimentary rock types so characteristic of both the slate belt and Avalon sequences, has so far yielded none of the late Precambrian fossils found in the others, and has so far yielded no radiometric dates as old as those from the Carolina slate belt (*see* Glover and Sinha, 1973). Moreover, we suspect that the James Run volcanic rocks were erupted above a west-dipping Benioff zone for the reasons already outlined. Perhaps there were two arcs facing one another, one which produced the James Run, and one which produced the Carolina slate belt. If so, the southern Piedmont could contain both, one along the western margin of the Inner Piedmont, and one in the eastern part of the Charlotte belt and in the Carolina slate belt.

REFERENCES CITED

BAILEY, E. B., and MACKIN, J. H., 1937, Recumbent folding in the Pennsylvania Piedmont; preliminary statement. Am. Jour. Sci., 5th Series, v. 33, p. 187-190.

BAKER, B. H., MOHR, P. A., and WILLIAMS, L. A. J., 1972, Geology of the eastern rift system of Africa. Geol. Soc. America Special Paper 136, 67 p.

BASCOM, F., and STOSE, G. W., 1932, Description of the Coatesville and West Chester Quadrangles (Pa.-Del.). U.S. Geol. Survey Geol. Atlas, Folio 223.

BIRD, J. M., and DEWEY, J. F., 1970, Lithosphere late-continental margin tectonics and the evolution of the Appalachian orogen. Geol. Soc. America Bull. v. 81, p. 1031-1060.

BROEDEL, C. H., 1937, The structure of the gneiss domes near Baltimore, Maryland, *in* Volume 13, Md. Geol. Survey, p. 149-188.

BROMERY, R. W., 1960, Preliminary interpretation of aeromagnetic data in the Allentown Quadrangle, Pennsylvania. U.S. Geol. Survey Prof. Paper 400-B, p. B178-B180.

——, 1968, Geological interpretation of aeromagnetic and gravity surveys of the northeastern end of the Baltimore-Washington anticlinorium, Harford, Baltimore, and part of Carroll County, Maryland. Unpublished Ph.D. dissertation, Johns Hopkins Univ., Baltimore, Md., 124 p.

——, and GRISCOM, A., 1967, Aeromagnetic and generalized geologic map of southeastern Pennsylvania. U.S. Geol. Survey Geophysical Investigations Map GP-577, scale 1:125,000.

BROUGHTON, J. G., FISHER, D. W. ISACHSEN, Y., and RICHARD, L. V., 1962, Geologic map of New York. The Univ. of the State of New York, State Education Dept., Map and Chart Series No. 5, 1:250,000.

BURKE, K., and DEWEY, J. F., 1973, Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. Jour. Geology, v. 81, p. 406-433.

CAMPOS, C. W. M., PONTE, F. C., and MIURA, K., 1974, Geology of the Brazilian continental margin, *in* Burk, C. A., and Drake, C. L., eds., The geology of continental margins. New York, Springer-Verlag, p. 447-462.

CHOQUETTE, P. W., 1960, Petrology and structure of the Cockeysville Formation (pre-Silurian) near Baltimore, Maryland. Geol. Soc. America Bull., v. 71, p. 1027-1052, 3 pls.

CLEAVES, E. T , EDWARDS, J., JR., GLASER, J. D., Compilers and editors, 1968, Geologic map of Maryland. Md. Geol. Survey, scale 1:250,000.

CLOOS, E., and HIETANEN, A., 1941, Geology of the "Martic overthrust" and the Glenarm Series in Pennsylvania and Maryland. Geol. Soc. America Special Paper 35, 207 p.

CONLEY, J. F., and HENIKA, W. S., 1973, Geology of the Snow Creek, Martinsville East, Price, and Spray Quadrangles, Virginia. Va. Div. of Mineral Resources, Rept. of Investigations 33, 71 p.

CROWLEY, W. P., 1976, The geology of the crystalline rocks near Baltimore and its bearing on the evolution of the eastern Maryland Piedmont. Md. Geol. Survey Rept. of Investigations No. 27, 40 p.

DAHLSTROM, C. D. A., 1970, Structural geology in the eastern margin of the Canadian Rocky Mountains. Bull. Canadian Petroleum Geology, v. 18, p. 332-406.

DALLMEYER, R. D , 1974, Tectonic setting of the northeastern Reading Prong. Geol. Soc. America Bull., v. 85, p. 131-134.

DEWEY, J. F., 1969, Evolution of the Appalachian/Caledonian orogen. Nature, v. 222, p. 124-129.

——, and BURKE, K., 1974, Hot spots and continental breakup: implications for collisional orogeny. Geology, v. 2, p. 57-60

DICKINSON, W. R., 1970, Relations of andesites, granites, and derivative sandstones to arc-trench tectonics. Rev. Geophysics and Space Physics, v. 8, p. 813-860.

DOELL, R. R., and COX, A., 1965, Measurement of the remanent magnetization of igneous rocks. U.S. Geol. Survey Bull. 1203-A, p. A1-A32.

DRAKE, A. A., JR., 1969, Precambrian and Lower Paleozoic geology of the Delaware Valley, New Jersey-Pennsylvania, *in* Subitzky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions. Rutgers Univ. Press, New Brunswick, N.J., p. 51-131.

——, 1970, Structural geology of the Reading Prong, *in*, Studies of Appalachian geology; central and southern, Fisher, G. W., and others, eds., Interscience, New York, p. 271-291.

FABIANO, E. B., and PEDDIE, N. W., 1969, Grid values of total magnetic intensity, IGRF, 1965. ESSA, Coast and Geodetic Survey Tech. Rept. 38, 55 p.

FISHER, G. W., 1963, The petrology and structure of the crystalline rocks along the Potomac River, near Washington, D.C. Unpublished Ph.D. dissertation, Johns Hopkins Univ., 241 p.

——, 1970, The metamorphosed sedimentary rocks along the Potomac River near Washington, D.C., *in* Studies of Appalachian geology; central and southern, Fisher, G. W., and others, eds., Interscience, New York, p. 299-316.

——, 1971, Kyanite-, staurolite-, and garnet-bearing schists in the Setters Formation, Maryland Piedmont. Geol. Soc. America Bull., v. 82, p. 229-232.

——, 1978, Geologic map of the New Windsor Quadrangle, Maryland. U.S. Geol. Survey, Misc. Inv. Series Map I-1037, scale 1:24,000.

FREEDMAN, J., WISE, D. U., and BENTLEY, R., 1964, Pattern of folded folds in the Appalachian Piedmont along Susquehanna River. Geol. Soc. America Bull., v. 75, p. 621-638.

GLOVER, L., and SINHA, A. K., 1973, The Virginia deformation, a Late Precambrian to Early Cambrian (?) orogenic event in the central Piedmont of Virginia and North Carolina. Am. Jour. Sci., v. 273A, p. 234-251.

GOHN, G. S., and THOMPSON, A. M., 1974, Lithofacies in the Marburg and Wissahickon Formations near the Martic Line, York and Lancaster Counties, Pennsylvania (abs.). Geol. Soc. America Abstracts with Programs, v. 6, no. 1, p. 31.

GRAY, C., and others, 1960, Geologic map of Pennsylvania. Pa. Geol. Survey, 4th Series, scale 1:250,000.

HADLEY, J. B., 1970, The Ocoee Series and its possible correlatives, *in* Studies of Appalachian geology; central and southern, Fisher, G. W., and others, eds., Interscience, New York, p. 247-260.

HALL, L. M., 1968, Times of origin and deformation of bedrock in the Manhattan Prong, *in* Studies of Appalachian geology, northern and maritime. Zen, E-an, and others, eds., Interscience, New York, p. 117-128.

HIGGINS, M. W., 1972, Age, origin, regional relations and nomenclature of the Glenarm Series, central Appalachian Piedmont: A reinterpretation. Geol. Soc. America Bull., v. 83, p. 984-1026.

——, 1973, Superimposition of folding in the northeastern Maryland Piedmont and its bearing on the history and tectonics of the central Appalachians. Am. Jour. Sci., v. 273A, p. 150-195.

——, and FISHER, G. W., 1971, A further revision of the stratigraphic nomenclature of the Wissahickon Formation in Maryland. Geol. Soc. America Bull., v. 82, p. 769-774.

——, FISHER, G. W., and ZIETZ, I., 1973, Aeromagnetic discovery of a Baltimore Gneiss dome in the Piedmont of northwestern Delaware and southeastern Pennsylvania. Geology, v. 1, p. 41-43.

——, SINHA, A. K., ZARTMAN, E. E., and KIRK, W. S., 1977, U-Pb zircon dates from the central Appalachian Piedmont: a possible case of inherited radiogenic lead. Geol. Soc. America Bull., v. 88, p. 125-132.

HOPSON, C. A., 1964, The crystalline rocks of Howard and Montgomery Counties; *in* The geology of Howard and Montgomery Counties, Md. Geol. Survey, p. 27-215.

JOHNSON, M. E., 1950, Geologic map of New Jersey. New Jersey State Dept. of Conservation and Development, scale 1:250,000.

KAY, G. M., 1941, Taconic allochthon and the Martic thrust. Science, 2nd Series, v. 94, p. 73.

KNOPF, E. B., and JONAS, A. I., 1929, The geology of the crystalline rocks of Baltimore County; *in* Baltimore County. Md. Geol. Survey, p. 97-199.

MACKIN, J. H., 1950, The downstructure method of viewing geologic maps. Jour. Geology, v. 58, p. 55-72.

——, 1962, Structure of the Glenarm Series in Chester County, Pennsylvania. Geol. Soc. America Bull., v. 73, p. 403-410.

McKINSTRY, H. E., 1961, Structure of the Glenarm Series in Chester County, Pennsylvania. Geol. Soc. America Bull., v. 72, p. 557-578.

MILLER, B. L., 1935, Age of the schists in the South Valley Hills, Pennsylvania. Geol. Soc. America Bull., v. 46, p. 715-756.

ODOM, A. L., and FULLAGAR, P. D., 1973, Geochronologic and tectonic relationships between the Inner Piedmont, Brevard zone, and Blue Ridge belts, North Carolina. Am. Jour. Sci., v. 273A, p. 133-149.

OLDENBURG, D. W., and BRUNE, J. N., 1975, An explanation for the orthogonality of ocean ridges and transform faults: Jour. Geophys. Research, v. 80, p. 2575-2585.

OLSEN, S. N., 1972, Petrology of the Baltimore Gneiss. Unpublished Ph.D. dissertation, Johns Hopkins Univ., Baltimore, Md., 295 p.

PETTIJOHN, F. J., 1975, Sedimentary rocks, 3rd ed., Harper and Row, New York, 628 p.

PIRSON, S. J., 1940, Polar charts for interpreting magnetic anomalies. Am. Inst. Mining and Metal. Engineers Trans., v. 138, p. 173-192.

RANKIN, D. W., 1975, The continental margin of eastern North America in the southern Appalachians: The opening and closing of the proto-Atlantic Ocean. Am. Jour. Sci., v. 275A, p. 298-336.

——, 1976, Appalachian salients and recesses: Late Precambrian continental breakup and the opening of Iapetus Ocean. Jour. Geophys. Research, v. 81, p. 5605-5619.

——, STERN, T. W., REED, J. C., JR. and NEWELL, M. F., 1969, Zircon ages of felsic volcanic rocks in the Upper Precambrian of the Blue Ridge Appalachian Mountains. Science, v. 166, No. 3906, p. 741-744.

RATCLIFFE, N. M., and KNOWLES, R. R., 1969, Stratigraphic relations along the western edge of the Cortlandt intrusives and their bearing on the Inwood-Manhattan problem. City Univ. of New York, Queens Coll. Dept. Geology, Geological Bull. 3, p. 49-64.

REINHARDT, J., 1974, Stratigraphy, sedimentology and Cambro-Ordovician paleogeography of the Frederick Valley, Maryland. Md. Geol. Survey Rept. of Investigations No. 23, 74 p.

RODGERS, J., 1968, The eastern edge of the North American continent during Cambrian and Early Ordovician, *in* Studies of Appalachian geology; northern and maritime; Zen, E-an, and others, eds., Interscience, New York, p. 141-149.

——, 1970, The tectonics of the Appalachians. Wiley-Interscience, New York, 271 p.

——, 1972, Latest Precambrian (post-Grenville) rocks of the Appalachian region. Am. Jour. Sci., v. 272, p. 507-520.

SCHWAB, F. L., 1972, The Chilhowee Group and the Late Precambrian-Early Paleozoic sedimentary framework in the central and southern Appalachians, *in* Appalachian structures: origin, evolution, and possible potential for new exploration frontiers, Lessing, P., and others, eds., West Va. Univ. and West Va. Geological and Economic Survey, p. 59-86.

SEIDERS, V. M., MIXON, R. B., STERN, T. W., NEWELL, M. F., and THOMAS, C. B., JR., 1975, Age of plutonism and tectonism and a new minimum age limit on the Glenarm Series in the northeast Virginia Piedmont, near Occoquan. Am. Jour. Sci., v. 275, p. 481-511.

SOUTHWICK, D. L., 1969, Crystalline rocks of Harford County. *in* The geology of Harford County, Maryland, Md. Geol. Survey, p. 1-76.

——, and OWENS, J. P., 1968, Geologic map of Harford County. Md. Geol. Survey, scale 1:62,500.

——, and FISHER, G. W., 1967, Revision of stratigraphic nomenclature of the Glenarm Series in Maryland. Md. Geol. Survey Rept. of Investigations No. 6, 19 p.

STOSE, A. J., and STOSE, G. W., 1944, Geology of the Hanover-York district, Pennsylvania. U.S. Geol. Survey Prof. Paper 204, 84 p.

——, 1946, Geology of Carroll and Frederick Counties, *in* The physical features of Carroll and Frederick Counties, Md. Geol. Survey, p. 11-131.

STOSE, G. W., and JONAS, A. I., 1939, Geology and mineral resources of York County, Pennsylvania. Pa. Topographic and Geologic Survey, County Report G7, 199 p.

STOSE, G. W., and STOSE, A. J., 1948, Stratigraphy of the Arvonia Slate, Virginia. Am. Jour. Sci., v. 246, p. 393-412.

TILTON, G. R., DOE, B. R., and HOPSON, C. A., 1970, Zircon age measurements in the Maryland Piedmont, with special reference to Baltimore Gneiss problems, *in* Studies in Appalachian geology, central and southern. Fisher, G. W., and others, eds., New York, Interscience, p. 429-437.

U.S. GEOLOGICAL SURVEY, 1966, Geological Survey Research, 1966, U.S. Geol. Survey Prof. Paper 550-A, 385 p.

WAGNER, M. E., and CRAWFORD, M. L., 1975, Polymetamorphism of the Precambrian Baltimore Gneiss in southeastern Pennsylvania. Am. Jour. Sci., v. 275, p. 653-682.

WALCOTT, C. D., 1892, Notes on the Cambrian rocks of Virginia and the southern Appalachians. Am. Jour. Sci., 3rd Series, v. 44, p. 52-57.

WISE, D. U., 1970, Multiple deformation, geosynclinal transitions and the mantle problem in Pennsylvania, *in* Studies of Appalachian geology: central and southern. Fisher, G. W., and others, New York, Interscience, p. 317-334.

MARYLAND GEOLOGICAL SURVEY
Merryman Hall
The Johns Hopkins University
Baltimore, Maryland 21218

Reports of Investigations:

1. Chemical quality of water and trace elements in the Patuxent River Basin, by S. G. Heidel and W. W. Frenier, 1965, 40 p. $3.50*
2. Pleistocene stratigraphy of the Salisbury area, Maryland and its relationship to the lower Eastern Shore: a subsurface approach, by H. J. Hansen, 1966, 56 p. 2.25
3. Water resources of the Salisbury area, Maryland, by D. H. Boggess and S. G. Heidel, 1968, 69 p. 3.00
4. Expandable clay in the St. Mary's Formation of Southern Maryland, by M. M. Knechtel, H. P. Hamlin, and J. W. Hosterman, 1966, 19 p. ... 1.00
5. Chemical quality reconnaissance of water of Maryland streams, by J. D. Thomas, 1966, 61 p. ... 2.00
6. Revision of stratigraphic nomenclature—Glenarm Series of the Appalachian Piedmont, by D. L. Southwick and G. W. Fisher, 1967, 19 p. 1.50
7. Geophysical log cross-section network of the Cretaceous sediments of Southern Maryland, by H. J. Hansen, 1968, 56 p 4.00
8. Piedmont and Coastal Plain geology along the Susquehanna Aqueduct: Baltimore to Aberdeen, Maryland, by E. T. Cleaves, 1968, 45 p. 2.00
9. Chemical and physical character of municipal water supplies in Maryland, by J. D. Thomas and S. G. Heidel, 1969, 52 p. 1.00
10. Ground-water occurrence in the Maryland Piedmont, by L. J. Nutter and E. G. Otton, 1969, 56 p. 2.50
11. Petrology and origin of Potomac and Magothy (Cretaceous) sediments, Middle Atlantic Coastal Plain, by J. D. Glaser, 1969, 101 p. 3.50
12. Paleoecology of the Choptank Formation (Miocene) of Maryland and Virginia, by R. E. Gernant, 1970, 90 p. 4.50
13. Extent of brackish water in the tidal rivers of Maryland, by W. E. Webb and S. G. Heidel, 1970, 46 p. 1.50
14. Geologic and hydrologic factors bearing on subsurface storage of liquid wastes in Maryland, by E. G. Otton, 1970, 39 p. 2.75
15. Geology and mineral resources of Southern Maryland, by J. D. Glaser, 1971, 84 p. 4.75
16. Flow characteristics of Maryland streams, by P. N. Walker, 1971, 160 p. 2.50
17. Water resources of Dorchester and Talbot Counties, Maryland with special emphasis on the groundwater potential of the Cambridge and Easton areas, by F. K. Mack, W. E. Webb, and R. A. Gardner, 1971, 107 p. 5.25
18. Solid-waste disposal in the geohydrologic environment of Maryland, by E. G. Otton, 1972, 59 p. 3.00
19. Hydrogeology of the carbonate rocks, Frederick and Hagerstown Valleys, Maryland, by L. J. Nutter, 1973, 70 p. 3.50
20. Hydrogeology of the formation and neutralization of acid waters draining from underground mines of western Maryland, by E. F. Hollyday and S. W. McKenzie, 1973, 50 p. 2.25
21. Sedimentary facies of the Aquia Formation in the subsurface of the Maryland Coastal Plain, by H. J. Hansen, 1974, 47 p. 3.50
22. An evaluation of the Magothy aquifer in the Annapolis area, Maryland, by R. K. Mack, 1974, 75 p. 3.00
23. Stratigraphy, sedimentology, and Cambro-Ordovician paleogeography of the Frederick Valley, Maryland, by J. Reinhardt, 1974, 74 p. 4.50
24. Availability of fresh ground water in northeastern Worcester County, Maryland: with special emphasis on the Ocean City area, by J. M. Weigle, 1974, 64 p. 4.00
25. Petrologic and chemical investigation of chemical weathering in mafic rocks, eastern Piedmont of Maryland, by E. T. Cleaves, 1974, 28 p. 2.25
26. Hydrogeology of the Triassic rocks of Maryland, by L. J. Nutter, 1975, 37 p. 2.50
27. The geology of the crystalline rocks near Baltimore bearing on the evolution of the eastern Maryland Piedmont, by W. P. Crowley, 1976, 40 p. 4.75
28. Digital simulation and prediction of water levels in the Magothy aquifer in Southern Maryland, by F. K. Mack and R. J. Mandle, 1977, 42 p. 3.00
29. Upper Cretaceous (Senonian) and Paleocene (Danian) pinchouts on the south flank of the Salisbury Embayment, Maryland, and their relationship to antecedent basement structures, by H. J. Hansen, 1978, 36 p. 5.25
30. New data bearing on the structural significance of the upper Chesapeake Bay magnetic anomaly, by J. Edwards, Jr. and H. J. Hansen, 1979, 44 p. 3.75
31. Simulated changes in water level in the Piney Point aquifer in Maryland, by J. Williams, 1979, 50 p. 6.25
32. Geological interpretations of aeromagnetic maps of the crystalline rocks of the central Appalachians, northern Virginia to New Jersey, by G. W. Fisher, M. W. Higgins, and I. Zietz, 1979, 43 p. 6.25
33. A quasi three-dimensional finite-difference groundwater flow model with a field application, by Grufron Achmad, 1979 ..... 6.50

* microfiche

Maryland Residents Add 5% Sales Tax

Aeromagnetic map of eastern Pennsylvania and northern New Jersey showing generalized geologic relations of the crystalline rocks.

GEORGE W. FISHER, MICHAEL W. HIGGINS AND ISODORE ZIETZ.

EXPLANATION

Magnetic intensity in gammas

1600
1400
1200
1000
800
600
400
200

JTR — Newark Group (Jurassic and Triassic) and related igneous intrusions

Unconformity

D — Devonian sedimentary rocks

S — Silurian sedimentary rocks

OC — Ordovician and Cambrian sedimentary rocks

Unconformity

pC — Precambrian rocks (includes minor inliers of OC in southwestern part of Reading Prong)

Structure symbols

Contact, dotted where concealed

High-angle fault

REFERENCES CITED

Bromery, R. W., 1967, Aeromagnetic map of Baltimore County and Baltimore City, Maryland: U. S. Geol. Survey Geophys. Inv. Map GP-613, scale 1:62,500.

Bromery, R. W., Petty, A. J., and Smith, C. W., 1964, Aeromagnetic map of Bel Air and vicinity, Harford, Baltimore, and Cecil Counties, Maryland: U. S. Geol. Survey Geophys. Inv. Map GP-482, scale 1:62,500.

Bromery, R. W., and Griscom, Andrew, 1967, Aeromagnetic and generalized geologic map of southeastern Pennsylvania: U. S. Geol. Survey Geophys. Inv. Map GP-577, scale 1:125,000.

Fabiano, E. B., and Peddie, N. W., 1969, Grid values of total magnetic intensity, IGRF, 1965: ESSA, Coast and Geodetic Survey Tech. Rept. 38, 55 p.

Henderson, J. R., Andreasen, G. E., and Petty, A. J., 1966, Aeromagnetic map of northern New Jersey and adjacent parts of New York and Pennsylvania: U. S. Geol. Survey Geophys. Inv. Map GP-562, scale 1:125,000.

Henderson, J. R., Johnson, R. W., and Gilbert, F. P., 1963, Aeromagnetic map of the Wilmington, Delaware, area and adjacent parts of Pennsylvania and Maryland: U. S. Geol. Survey Geophys. Inv. Map GP-363, scale 1:62,500.

U. S. Geological Survey, 1971a, Aeromagnetic map of part of Kent County, Maryland, and parts of adjacent counties in Maryland and Delaware: U. S. Geol. Survey Geophys. Inv. Map GP-756, scale 1:62,500.

U. S. Geological Survey, 1971b, Aeromagnetic map of part of Cecil County, Maryland, and parts of adjacent counties in Maryland and Delaware: U. S. Geol. Survey Geophys. Inv. Map GP-755, scale 1:62,500.

U. S. Geological Survey, 1973, Aeromagnetic map of northern Frederick County, Maryland, and parts of adjacent counties in Maryland and Pennsylvania: U. S. Geol. Survey open-file map, scale 1:62,500.

U. S. Geological Survey, 1974a, Aeromagnetic map of Carroll County, Maryland: U. S. Geol. Survey open-file map 74-205, scale 1:62,500.

U. S. Geological Survey, 1974b, Aeromagnetic map of Frederick County, Maryland: U. S. Geol. Survey open-file map 74-206, scale 1:62,500.

U. S. Geological Survey, 1974c, Aeromagnetic map of Howard County, Maryland: U. S. Geol. Survey open-file map 74-207, scale 1:62,500.

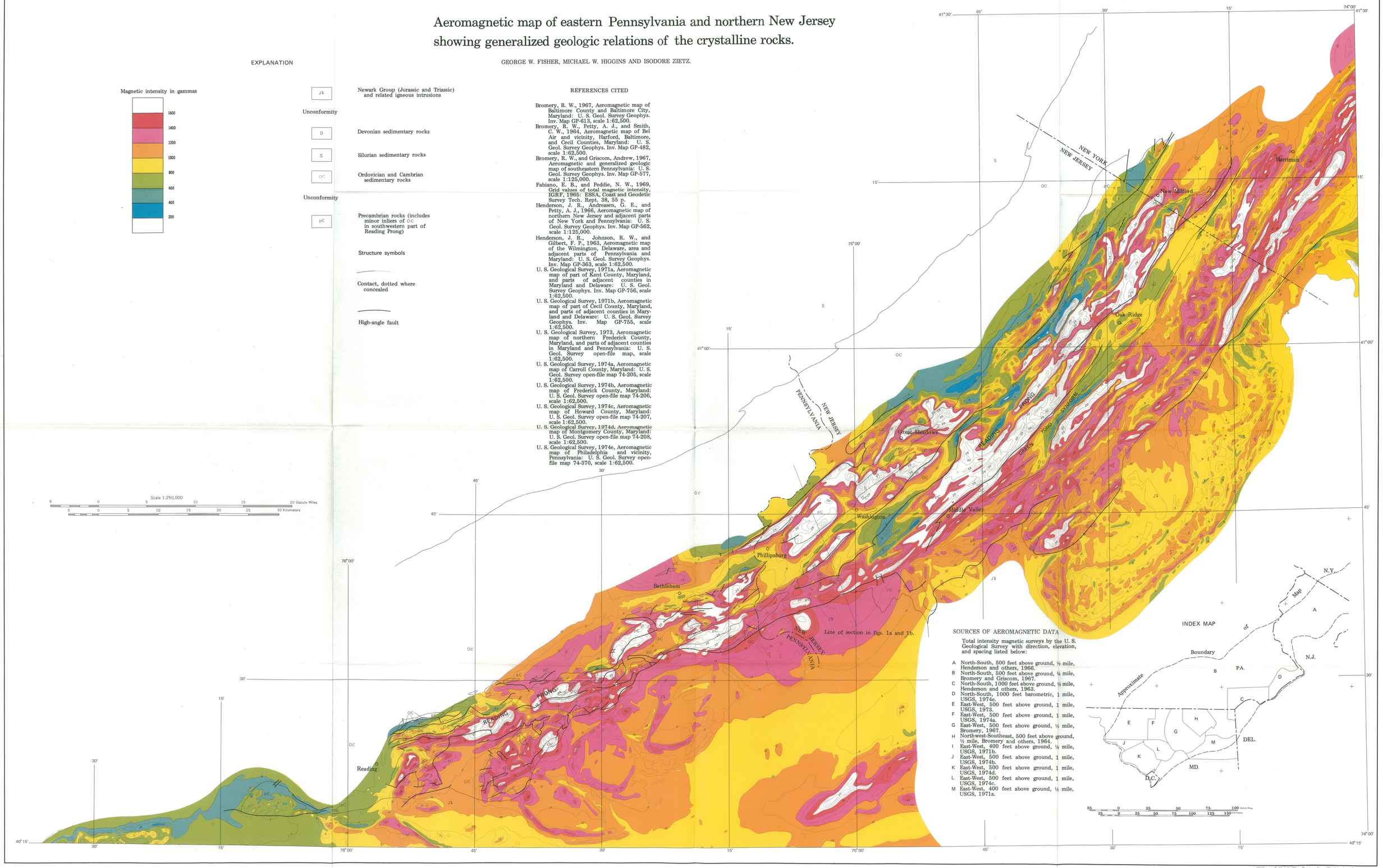
U. S. Geological Survey, 1974d, Aeromagnetic map of Montgomery County, Maryland: U. S. Geol. Survey open-file map 74-208, scale 1:62,500.

U. S. Geological Survey, 1974e, Aeromagnetic map of Philadelphia and vicinity, Pennsylvania: U. S. Geol. Survey open-file map 74-370, scale 1:62,500.

Scale 1:250,000

5 0 5 10 15 20 Statute Miles

5 0 5 10 15 20 25 30 Kilometers

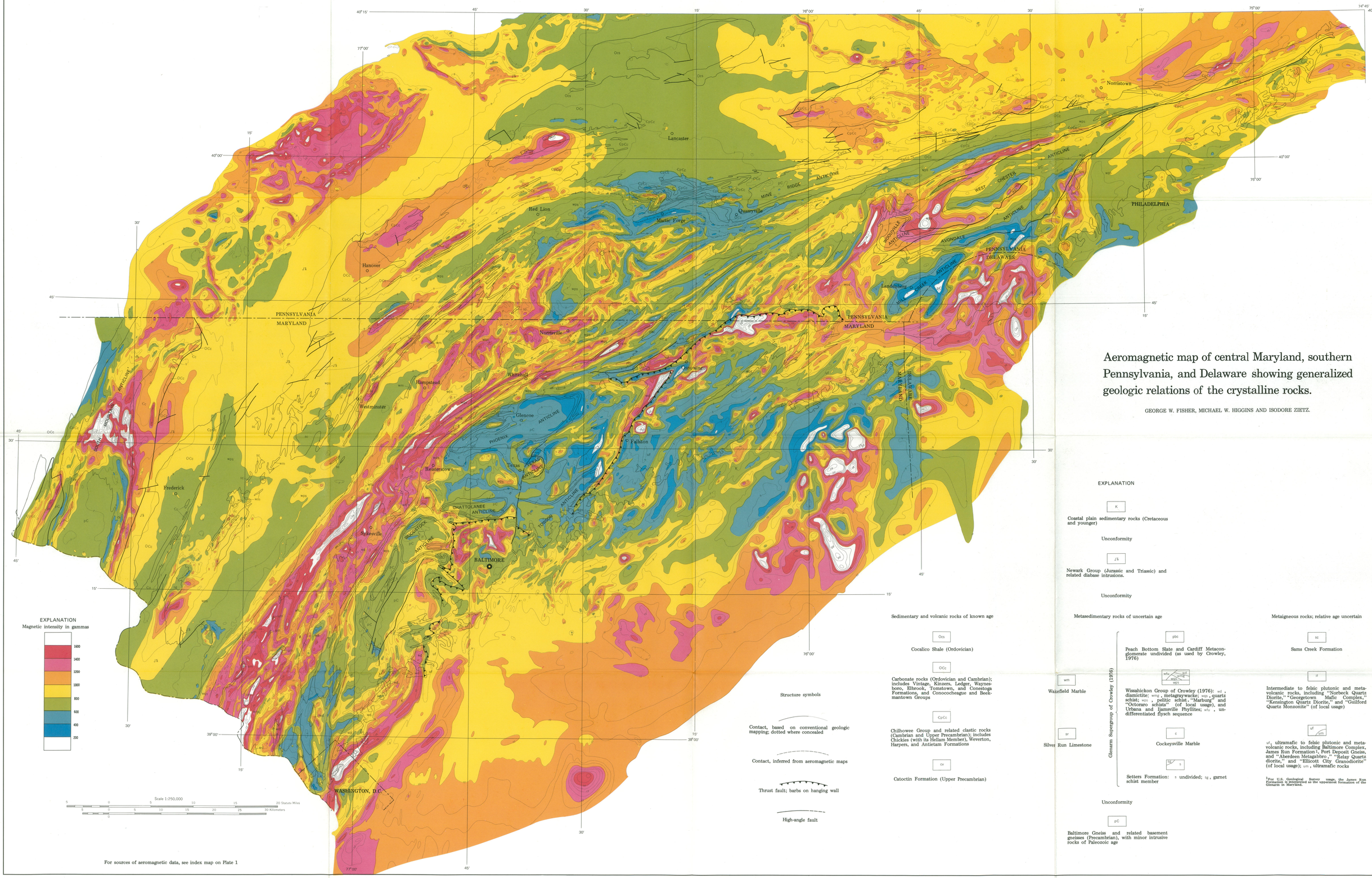


SOURCES OF AEROMAGNETIC DATA

Total intensity magnetic surveys by the U. S. Geological Survey with direction, elevation, and spacing listed below:

A North-South, 500 feet above ground, ½ mile, Henderson and others, 1966.
B North-South, 500 feet above ground, ¼ mile, Bromery and Griscom, 1967.
C North-South, 1000 feet above ground, ½ mile, Henderson and others, 1963.
D North-South, 1000 feet barometric, 1 mile, USGS, 1974e.
E East-West, 500 feet above ground, 1 mile, USGS, 1973.
F East-West, 500 feet above ground, 1 mile, USGS, 1974a.
G East-West, 500 feet above ground, ½ mile, Bromery, 1967.
H Northwest-Southeast, 500 feet above ground, ½ mile, Bromery and others, 1964.
I East-West, 400 feet above ground, ½ mile, USGS, 1971b.
J East-West, 500 feet above ground, 1 mile, USGS, 1974b.
K East-West, 500 feet above ground, 1 mile, USGS, 1974d.
L East-West, 500 feet above ground, 1 mile, USGS, 1974c.
M East-West, 400 feet above ground, ½ mile, USGS, 1971a.

Williams & Heintz Map Corporation, Washington, D.C. 20027



For sources of aeromagnetic data, see index map on Plate 1

Aeromagnetic map of central Maryland, southern Pennsylvania, and Delaware showing generalized geologic relations of the crystalline rocks.

GEORGE W. FISHER, MICHAEL W. HIGGINS AND ISODORE ZIETZ.

EXPLANATION

K

Coastal plain sedimentary rocks (Cretaceous and younger)

Unconformity

J'k

Newark Group (Jurassic and Triassic) and related diabase intrusions.

Unconformity

Sedimentary and volcanic rocks of known age

Ocs

Cocalico Shale (Ordovician)

OCc

Carbonate rocks (Ordovician and Cambrian); includes Vintage, Kinzers, Ledger, Waynesboro, Elbrook, Tomstown, and Conestoga Formations, and Conococheague and Beekmantown Groups

CpCc

Chilhowee Group and related clastic rocks (Cambrian and Upper Precambrian); includes Chickies (with its Hellam Member), Weverton, Harpers, and Antietam Formations

cv

Catoctin Formation (Upper Precambrian)

Metasedimentary rocks of uncertain age

wm

Wakefield Marble

sr

Silver Run Limestone

Glenarm Supergroup of Crowley (1976)

pbc

Peach Bottom Slate and Cardiff Metaconglomerate undivided (as used by Crowley, 1976)

wfu, wd, wmg, wgs

Wissahickon Group of Crowley (1976): wd, diamictite; wmg, metagraywacke; wgs, quartz schist; wgs, pelitic schist, "Marburg" and "Octoraro schists" (of local usage), and Urbana and Ijamsville Phyllites; wfu, undifferentiated flysch sequence

c

Cockeysville Marble

sg, s

Setters Formation: s undivided; sg, garnet schist member

Metaigneous rocks; relative age uncertain

sc

Sams Creek Formation

if

Intermediate to felsic plutonic and metavolcanic rocks, including "Norbeck Quartz Diorite," "Georgetown Mafic Complex," "Kensington Quartz Diorite," and "Guilford Quartz Monzonite" (of local usage)

uf, um

uf, ultramafic to felsic plutonic and metavolcanic rocks, including Baltimore Complex, James Run Formation[1], Port Deposit Gneiss, and "Aberdeen Metagabbro," "Relay Quartz diorite," and "Ellicott City Granodiorite" (of local usage); um, ultramafic rocks

[1]For U.S. Geological Survey usage, the James Run Formation is interpreted as the uppermost formation of the Glenarm in Maryland.

Unconformity

pC

Baltimore Gneiss and related basement gneisses (Precambrian), with minor intrusive rocks of Paleozoic age

Structure symbols

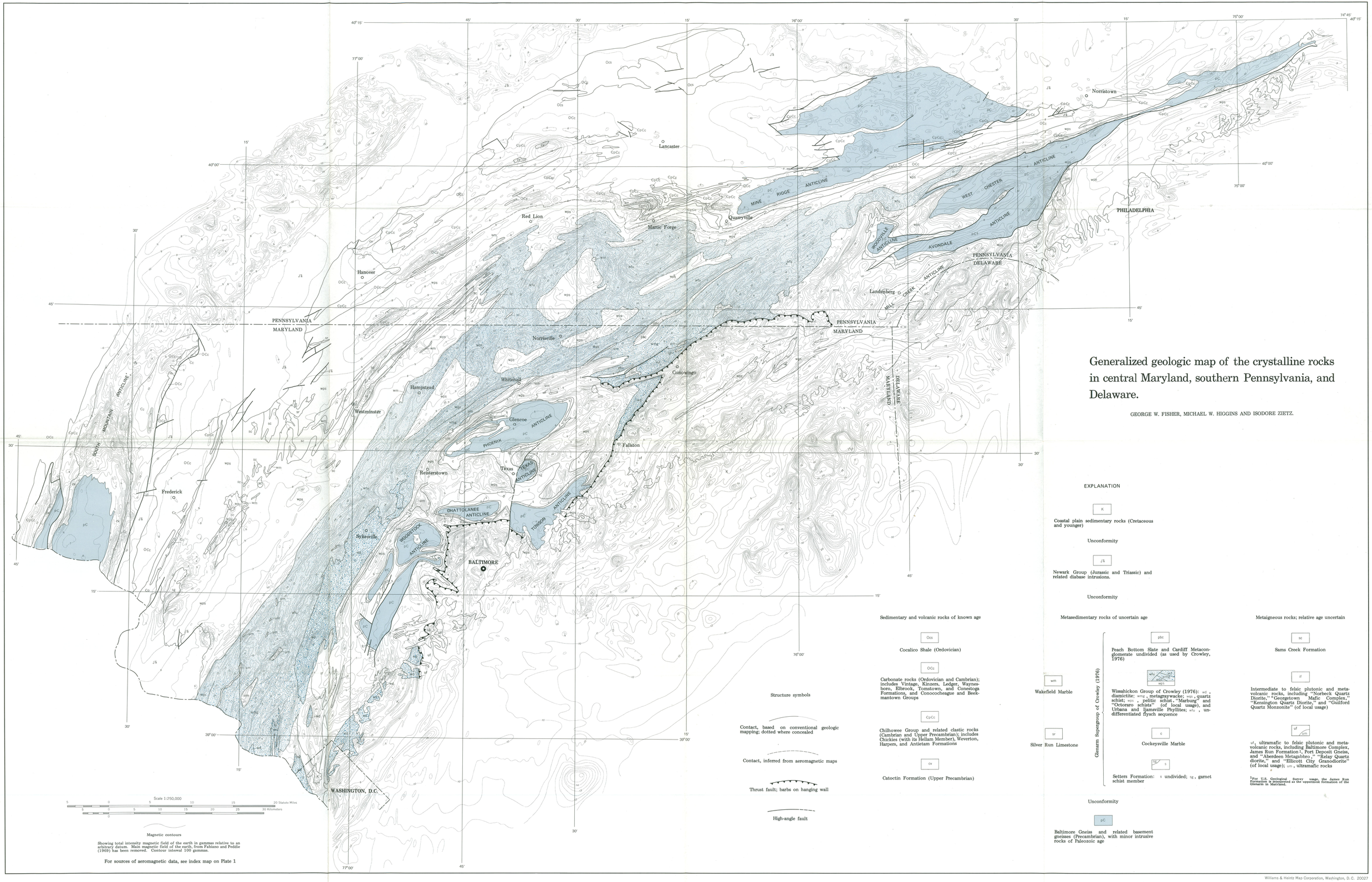
Contact, based on conventional geologic mapping; dotted where concealed

Contact, inferred from aeromagnetic maps

Thrust fault; barbs on hanging wall

High-angle fault

Williams & Heintz Map Corporation, Washington, D.C. 20027



Generalized geologic map of the crystalline rocks in central Maryland, southern Pennsylvania, and Delaware.

GEORGE W. FISHER, MICHAEL W. HIGGINS AND ISODORE ZIETZ.

Williams & Heintz Map Corporation, Washington, D.C. 20027

E 2

5'

GEOLOGICAL INTERPRETATION OF AEROMAGNETIC MAPS OF THE CRYSTALLINE ROCKS IN THE CENTRAL APPALACHIANS, NORTHERN VIRGINIA TO NEW JERSEY

R.I. No. 32, 1979